GR focus review

Middle–Late Paleozoic Australia–Asia convergence and tectonic extrusion of Australia

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ARTICLE INFO

Article history:

Received 5 January 2012

Received in revised form 3 October 2012

Accepted 13 October 2012

Available online 8 November 2012

Keywords:

Alice Springs Orogeny

Central Asian Orogenic Belt

Kiaman

paleomagnetism

tectonic extrusion

Variscan Orogeny

ABSTRACT

Recent and new paleomagnetic data from ignimbrite-rich Carboniferous successions of the western Tamworth Belt, Southern New England Orogen, eastern Australia, show a northward excursion over $\sim 30^\circ$. Paleozoic data from the Australian craton and Tasman Orogenic System (TOS) suggest an Early Devonian start. At the middle-late Visean peak, the central New Guinean promontory of the Australian craton reached 30° – 40° N, within the latitude range of the western Central Asian Orogenic Belt (CAOB). Devonian–Carboniferous convergence of Australia/northeastern Gondwana with the CAOB, across the Paleoasian–Rheic Ocean, is proposed as a major driver for contemporaneous tectonism throughout Australia and the CAOB. This implies a substantial Variscan, Pangea-forming, influence on Australian Devonian–Carboniferous tectonics – Alice Springs Orogeny (ASO) and Quilpie and Kanimblan Orogenies. Convergence-related compressional deformation of Australia is largely confined to a “compression box”, extending southward from the New Guinean promontory and bounded westward by the Lasserter Shear Zone and eastward by the East Australian Rift System. Comparable characteristics of Paleozoic Australia–Asia and Cenozoic India–Asia convergence – north–south compression, weak and heated crust (Larapintine Graben and TOS/Tibetan Plateau), eastern “free oceanic boundary” (Paleopacific/Pacific) – do link Paleozoic Australia–Asia convergence to Cenozoic tectonic extrusion of Tibet. Tectonic extrusion of ductile lower crust from the Larapintine Graben led to eastward displacement of the Thomson and Northern New England Orogens, with upper crustal displacement bounded northward by Arunta Block shear zones, the Diamantina River Lineament, the Clarke River Fault Zone and the Townsville Trough, and southward by the Darling River/Cobar-Inglewood Lineaments and Cato Fracture Zone with the Lake Blanche-Olepoloko Fault Zones and Lachlan Transverse Zone as a subsidiary. Recognition of ASO-related tectonic extrusion opens novel, provocative, insights into puzzling aspects of Australian Middle–Late Paleozoic evolution.

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1. Introduction

1.1. Alice Springs Orogeny – an orogeny in search of a cause

The cause of the Alice Springs Orogeny (ASO) is unclear, as summarized by Coney et al. (1990, p 537): “It obviously records a massive intraplate failure of the Australian craton. In a broad sense the Alice Springs orogeny is coeval with the mid-Carboniferous Kanimblan late compressive events in the Lachlan,..... To date, no plate tectonic explanation for the Alice Springs orogeny has been recognized.”

Australia’s Paleozoic tectonic evolution is generally interpreted in a Paleopacific subduction framework. A westward subduction driver however, does not relate naturally to the north–south stress pattern of the ASO. Alternative exploration of a southward, Tethyan tectonic, driver for the ASO has been hold back by prominent presence on Paleozoic global reconstructions of an eastward open ocean between northeastern Gondwana and southeastern Laurasia. Implications from recent paleomagnetic data from eastern Australia however, do question the reality of such an eastward open ocean during Early Carboniferous in particular. The new data define a substantial and mainly Early Carboniferous northward movement of Australia/northeastern Gondwana, leading to collision with the Central Asian Orogenic Belt (CAOB). Such convergence offers a wanting Tethyan driver for the ASO and, in turn, offers a conceptual stepping stone for re-interpreting Australian Middle–Late Paleozoic tectonics as driven, at least partly, by ASO-related tectonic extrusion. Such re-interpretation is attempted herein, building on new concepts for lower crustal ductile flow that are emerging from study of tectonic extrusion of the Tibetan Plateau by the India–Asia collision – arguably the world’s most spectacular example of Tethyan-driven intra-cratonic deformation, quite maturely developed yet actively progressing before

our eyes, and attracting a broad range of “hot” studies that are sourcing new, fundamental, concepts.

1.2. Unconventional leads

Powell (1984) recognized the widespread nature of Late Paleozoic north–south compressional deformation throughout Australia and attributed intra-cratonic ASO-related deformation and large-scale mega-kinking throughout the Tasman Orogenic System (TOS) to a common, though unidentified, mid-Carboniferous cause. The wide extent and pervasiveness of such deformation are well recognized nowadays (Williams and Pulford, 2008), exemplified by recent seismic profiling evidence for Early Carboniferous southward thrusting of the Thomson Orogen over the Lachlan Orogen at the Olepoloko Fault Zone (Glen et al., 2006) and Devonian? southward thrusting at the northern Selwyn Block (Cayley et al., 2011). This emerging picture of Australia-wide north–south compressional deformation however, is hard to reconcile with the plate tectonic paradigm of westward, head-on or even highly oblique, Paleopacific subduction as the main driver for Paleozoic deformation of the TOS and the Australian craton (Gray and Foster, 2004).

Several unconventional, directionally more comfortable, tectonic scenarios do invoke Tethyan, rather than Pacific, drivers featuring northward convergence and collision of Gondwana with Laurasia. Veevers (1994) and Veevers et al. (1994a) linked Late Paleozoic deformation of Australia and mid-Carboniferous uplift of the Gamburtsev Subglacial Mountains of central Antarctica to the well-established, but distant, Pangea-forming collision of northwestern Gondwana with southwestern Laurasia. In contrast, Klootwijk (1995, 1996a) attributed the ASO to convergence of Australia, as part of northeastern Gondwana,

with southeastern Laurasia. Veevers' attribution of eastern Gondwanan deformation to western Pangean collision followed the plate tectonic paradigm of a wide eastward open "Paleotethys" during Middle and Late Paleozoic (Fig. 1; Scotese, 2001; Golonka, 2002) and invoked transmission of far-field forces halfway across the globe, but offered no mechanism for focussing of deformation within easternmost Gondwana. Klotwijk's unconventional scenario featured convergence and collision of Australia/northeastern Gondwana with southeastern Laurasia based on, then preliminary, paleomagnetic evidence for a Middle–Late Paleozoic northward excursion of eastern Gondwana, proposed transmission of far-field forces over shorter distances as the cause of Middle–Late Paleozoic contemporaneous deformation of Australia and the Central Asian Orogenic Belt (CAOB), and so implied extension of the Variscan Orogeny concept to Australia.

This study outlines the Early Carboniferous northward excursion of Australia/northeastern Gondwana as currently defined from paleomagnetic data, proposes Middle–Late Paleozoic Australia–Asia convergence and closure of the Paleoasian–Rheic Ocean as the driver for ASO-related deformation, and explores tectonic extrusion of central and eastern Australia, guided by an emerging model for tectonic extrusion of Tibet driven by ongoing Cenozoic India–Asia convergence upon closure of the Neotethys.

2. Convergence across the Paleoasian–Rheic ocean

2.1. An eastward open "Paleotethys" before Late Carboniferous?

Prevalent global reconstructions for Carboniferous, Devonian and earlier show a wide eastward open ocean, often identified as "Paleotethys", between Gondwana and Laurasia (Fig. 1). Such reconstructions derive primarily from paleomagnetic pole paths for the two supercontinents. The Gondwana path largely reflects the prevailing Australian SLP (Schmidt, Li and Powell) path, with the KG (Klotwijk and Giddings) path proposed as an Australian alternative (see Klotwijk, 2009, 2010 and the references therein). The KG path

is based on paleomagnetic data for cratonic Australia and the TOS. Data from the New England Orogen (NEO) are excluded, yet the KG path resembles a detailed pole path based solely on data from the Southern New England Orogen (SNEO) (Klotwijk, 2002, 2003, 2009). The SLP, KG and SNEO pole paths have comparable shapes for latest Carboniferous and do support an eastward open ocean in eastern Pangea during Late Carboniferous. However, the SLP pole path diverges from the KG and SNEO paths for Early Carboniferous and Late Devonian, with the SLP path supporting an eastward open ocean and the KG and SNEO paths not.

Prevalent Pangea reconstructions rationalize Variscan deformation of the CAOB by northward convergence across the "Paleotethys" of peri-Gondwana-derived fragments such as Tarim and North China. Such convergence does not rationalize contemporaneous north–south deformation of Australia, nor does Paleopacific subduction. Convergence between Australia/northeastern Gondwana and the CAOB in contrast, provides a rational driver for Variscan deformation of the CAOB and contemporaneous north–south deformation of Australia. Extension of the Variscan Orogeny concept to Australia and northeastern Gondwana is novel and provocative and warrants close scrutiny of premises. What are the paleomagnetic evidence for Australia–Asia convergence and the geological evidence for contemporaneous deformation in Laurasia and Gondwana?

2.2. Contemporaneous tectonism in southern Laurasia and northeastern Gondwana

2.2.1. Southern Laurasia

The Variscan (~Hercynian) Orogeny concept was defined originally for western Europe (Kossmat, 1927; Torsvik and Cocks, 2009; Nance et al., 2010), but nowadays covers all Middle–Late Paleozoic (Late Devonian to Early Permian) deformation related to convergence of Gondwana with Laurasia to form Pangea. Such deformation extends over a global low-paleolatitude region, from the Ouachita–Appalachian fold belts of eastern North America and the conjugate

Fig. 1. Early Carboniferous Pangea reconstruction (Scotese, 2001). Legend: AF = Africa; AN = Antarctica; AU = Australia; AR = Arabia; BA = Baltica; IN = India; KA = "Kazakhstan"; LA = Laurentia; NC = North China; "PT/PA" = "Paleotethys/Paleoasian Ocean"; RH = former Rheic Ocean; SA = South America; SC = South China; SI = Siberia; TA = Tarim; Laurussia = LA + BA; Laurasia = LA + BA + SI + KA + TA + Cathaysia; Cathaysia = NC + SC + IC. See Section 2.4 for discussion of the Paleoasian–Rheic Ocean and Paleotethys concepts.

Mauritanide fold belts of northwestern Africa, across central Europe, the Balkan and Anatolia into the CAOB (Fig. 2; Winchester et al., 2002; Simancas et al., 2005; Nance and Linnemann, 2008; Biske and Seltmann, 2010; Michard et al., 2010).

2.2.2. Eastern Gondwana

It is argued herein that the Variscan Orogeny concept be extended to Devonian–Early Carboniferous ASO-related deformation of Australia (Powell, 1984; Veevers et al., 1994a; Li and Powell, 2001; McLaren et al., 2002; Williams and Pulford, 2008). East–west compressional deformation of the TOS, in the Kanimblan Orogen (Veevers, 1984; Scheibner and Veevers, 2000; Glen, 2005) and Quilpie Orogen (Finlayson, 1990a; Finlayson et al., 1990b; Leven et al., 1990), is commonly interpreted as driven by westward subduction of Paleopacific lithosphere beneath eastern Australia. However, the broadly north–south compressional regimes of the wider ASO and the TOS partly are hard to interpret in terms of Paleopacific subduction, even if highly oblique (Gray and Foster, 2004).

There is ample evidence of Devonian–Carboniferous N–S to NW–SE compression throughout Australia, demonstrated in particular for the central Australian Early–Middle Devonian Pertnjarra Movement and middle Carboniferous Mount Eclipse Movement (Fig. 3). Australia-wide extent of ASO-related deformation was highlighted by Powell and co-workers (Powell, 1984; Veevers, 1984; Powell et al., 1985) in relating mid–Carboniferous kink-banding throughout the eastern seaboard of the TOS (Goscombe et al., 1994; Rickard and Love, 2000) with ASO deformation in central Australia. Wide extent and longevity of ASO-related tectonism (400–330 + Ma, or extended ASO 450–330 + Ma) are well established from tectonic and geochronological studies in central Australia (Dunlap et al., 1995; Sandiford and Hand, 1998; Haines et al., 2001; Flottmann et al., 2004; Maidment et al., 2007), southern Australia (McLaren et al., 2002) and eastern Australia (Glen et al., 2006; Cayley et al., 2011). The ASO is now

recognized as the “biggest Paleozoic intracratonic folding and thrusting event to have affected the Australian craton” (Williams and Pulford, 2008, p 837).

Beyond mainland Australia there is some evidence for Devonian–Early Carboniferous tectonism in the New Guinean region of purported head-on collision with the CAOB (Visser and Hermes, 1962; Bain et al., 1975; Pigram and Panggabean, 1984; Milsom, 1985; Van Wyck and Williams, 2002). Part of the evidence relates however, to the Bird’s Head–Misool region, which may (Pigram and Davies, 1987), or may not (Dow and Sukanto, 1986), represent a westward displaced terrane. Speculative attribution of uplift of the Gamburtsev Subglacial Mountains of central Antarctica to mid–Carboniferous, Pangean-forming, compression (Veevers, 1994; Veevers et al., 2008) would imply a far southern extent of Devonian–Carboniferous tectonism throughout eastern Gondwana. What would drive such a major orogenic phase?

2.2.3. Widespread eastern Pangean tectonism incompatible with eastward open ocean

An eastward open “Paleotethys” prior to Late Carboniferous is difficult to reconcile with contemporaneous tectonism across the margins and hinterlands of southeastern Laurasia and northeastern Gondwana. Occam’s Razor points to Gondwana–Laurasia convergence and closure of the intervening ocean as a simple mechanism – a logical alternative to the conjoint paradigms of eastern Gondwanan tectonics driven by oblique westward subduction of the Paleopacific, and Laurasian tectonics driven by northward convergence of peri–Gondwana-derived blocks across a wide eastward open ocean.

Extensive paleomagnetic data from the SNEO demonstrate a northward excursion during Early Carboniferous, interpreted as northward convergence during Devonian–Early Carboniferous of eastern Gondwana with the CAOB of southern Laurasia. This implies that Pangea-forming convergence was not confined to northwestern

Fig. 2. Variscan (Hercynian), Cimmerian and Alpine orogenic belts in central and southern Eurasia (after Şengör, 1984, figs. 7, 8 and Van der Voo, 1993, fig. 7.1). Legend: Austr = Australia; CAOB = Central Asian Orogenic Belt; IC = Indo China; Tar = Tarim.

Fig. 3. NE-SW to NNE-SSW compressional deformation patterns during the Early–Middle Devonian Pertnjara phase (A) and the mid Carboniferous Mount Eclipse-Meda phase (B) of the Alice Springs Orogeny (after OZ SEEBASE™, 2005, pp. 113, 116). Legend: AN = Antarctica; AU = Australia; IN = Greater India.

Gondwana and southwestern Laurasia, but extended eastward into northeastern Gondwana and southern Laurasia. Paleomagnetic control on such convergence is discussed, based on new and recent data from the SNEO and the western CAOB.

2.3. Paleomagnetic control on Australia–Asia convergence

2.3.1. Carboniferous paleomagnetism of the Southern New England Orogen (SNEO)

Paleomagnetic data (Klootwijk, 2002, 2003, in preparation) from three tectono-stratigraphic blocks of the western Tamworth Belt, SNEO (Fig. 4; Rocky Creek, Werrie, Rouchel; Roberts and Geeve, 1999; Opdyke et al., 2000; Roberts et al., 2003, 2004, 2006) have been integrated into an extensive pole path spanning most of the Carboniferous (Fig. 5A; Supplementary Table 1). The data are derived mainly from ignimbrites, abundantly present throughout the Carboniferous successions, and less so from volcanoclastic deposits. Ignimbrites have been studied by preference, because of their potential for stable deuteric magnetization. Initial pilot studies demonstrated that many of the ignimbrites do retain a primary magnetization in spite of widespread overprinting (Klootwijk, 1996a). There is also extensive SHRIMP U–Pb zircon age control for the ignimbrites (Claoué-Long et al., 1995; Roberts et al., 1995b; Opdyke et al., 2000; Roberts et al., 2004, 2006), although early dating results may have uncorrectable errors of 1% to 3% from, discontinued, usage of an inhomogeneous calibration standard (SL13; Black et al., 2003; Roberts et al., 2003; Williams and Pulford, 2008).

The Carboniferous successions of the Rocky Creek, Werrie and Rouchel Blocks formed part of the Tamworth Belt forearc basin and derived mainly from a western continental volcanic arc – variously identified as Baldwin Volcanic Arc (Scheibner and Basden, 1998; Scheibner and Veivers, 2000), Keepit Arc (Leitch et al., 2010) or Currabubula Arc (McPhie, 1987; Glen, 2005; Jeon et al., 2012) – that evolved from Devonian andesitic volcanism to dacitic, and finally rhyolitic, Carboniferous volcanism (Roberts et al., 1991; Opdyke et al., 2000), ceasing by latest Carboniferous (~305 Ma, Roberts et al., 2006; Cawood et al., 2011a, 2011b; Jeon et al., 2012). The volcanic arc has not been located in outcrop. It may be westward overthrust by the Tamworth Belt during the Hunter–Bowen Phase of the Gondwanide Orogeny, or may be relocated southward with respect to the SNEO by sinistral strike-slip during latest Carboniferous–Early Permian

formation of the Bowen–Gunnedah–Sydney basin system and oroclinal bending of the SNEO (see Section 4.5.6) (Korsch and Harrington, 1987; Klootwijk, 2009; Cawood et al., 2011a,b).

The north–south tectonic grain of the western Tamworth Belt varies little across and between the three studied blocks. Comparison of paleomagnetic results has not indicated rotation between the blocks and results have been integrated, as observed, into a common pole path for the SNEO. The Carboniferous stratigraphy of Tamworth Belt blocks is characterized by a lower and an upper succession, separated by a middle Carboniferous disconformity that spans a “Namurian” hiatus of varying duration between late Viséan and early Westphalian (Opdyke et al., 2000; Roberts et al., 2003, 2004, 2006). SHRIMP U–Pb zircon ages show its time span to vary from about 20 to less than 15 million years (myr) in the Rouchel Block (Roberts et al., 2006), from less than 10 to about 4 myr in the Werrie Block (Roberts et al., 2006), and to less than 5 myr in the Rocky Creek Block (Roberts et al., 2003, 2004).

The SNEO pole path data indicate major paleolatitudinal shifts during Carboniferous (Fig. 5B; Supplementary Table 1). An Early Carboniferous northward movement over more than 30° preceded a Late Carboniferous southward movement to polar latitudes. The southward movement is well established paleomagnetically (Irving, 1966; Schmidt, 1988; Klootwijk, 1996a, 2009) and is supported by a reduction in faunal diversity, brachiopods in particular but also corals and polyzoans, interpreted as a shift into cooler, more southern, bioclimatic zones. Change from a highly diverse cosmopolitan assemblage to a less diverse Gondwanan assemblage occurs in the late Viséan *Marginirugus barringtonensis* Zone (Campbell and McKellar, 1969; Runnegar and Campbell, 1976; Roberts, 1981, 1987). Disappearance of oolites from the sedimentary record (Campbell and McKellar, 1969) and appearance of glaciation markers further support the southward movement (Powell and Veivers, 1987; Veivers and Powell, 1987; Isbell et al., 2012). The Early Carboniferous northward excursion is a more recent paleomagnetic observation that remains to be corroborated. Its interpretation may be strengthened by anticipated narrowing of the paleomagnetic data gap across the “Namurian” hiatus from studies in progress on the Gresford and Myall Blocks of the southern Tamworth Belt (see Klootwijk, 1996a, 1996b for reconnaissance data) and on the southern Rocky Creek Block.

The peak of the northward excursion (~330 Ma; Fig. 5B) and the middle-late Viséan, paroxysmal, phase of Variscan deformation do

Fig. 4. (A) Australian crustal mega-elements (after Shaw et al., 1996, fig. 5). Box refers to (B) Overview of the Southern New England Orogen (after Korsch and Harrington, 1987, fig. 2) with outlines of three paleomagnetically studied blocks in the western Tamworth Belt. Green = Devonian–Carboniferous accretionary wedge; Yellow = Devonian–Carboniferous forearc basin of the Tamworth Belt.

coincide in time, suggesting a causal relationship. The substantial latitudinal span of the northward excursion would imply exceptionally high speeds for the succeeding southward movement if original SHRIMP U–Pb zircon datings are applied. However, such unrealistic

high southward speeds reduce toward speeds known for India's northward movement prior to collision with southern Asia (~20 cm/yr: Klotwijk et al., 1992), if allowance is made for homogeneity problems with the SL13 calibration standard. When the paleomagnetic data gap across the “Namurian” hiatus is reduced by studies currently in progress, then movement speeds can be detailed through a SHRIMP re-dating program using appropriate calibration standards (Black et al., 2003). Consequent speeds of movement will then be more firmly based than at present. If still regarded excessive, emphasis should shift from questioning their reality as “tectonically possible” toward searching for possible mechanisms.

2.3.2. Middle–Late Paleozoic paleolatitudinal evolution of Australia

Provenance linkage between the Lachlan Orogen and the Gamilaroi Terrane, a precursor terrane of the Tamworth Belt, is commonly interpreted as accretion of the SNEO to the Lachlan Orogen by latest Devonian (Flood and Aitchison, 1992; Aitchison et al., 1999; Li and Powell, 2001; Champion et al., 2009; Glen and Roberts, 2010), although some do regard evidence equivocal till Late Carboniferous (Cawood and Leitch, 1985; Powell et al., 1990; Skilbeck and Cawood, 1994; McElhinny et al., 2003). The Calliope Arc of the NNEO accreted to the Thompson Orogen during Middle Devonian (Murray, 1986; Veevers et al., 1994b; Aitchison et al., 1999; Champion et al., 2009; Blake and Withnall, 2010), or more likely during Late Devonian (Murray et al., 2012). The Calliope Arc (NNEO) and the Gamilaroi Terrane (SNEO) have been interpreted as parts of the same intra-oceanic island arc (Aitchison et al., 1999; Offler and Gamble, 2002; Murray and Blake, 2005; Champion et al., 2009). Consensus has developed that this Calliope–Gamilaroi island arc accreted to the Thomson–Lachlan Orogen(s) by Middle–latest Devonian, with subsequent development of a continental arc, the Keepit–Connors Arc (Leitch et al., 2010; Cawood et al., 2011a,b), made up of the outcropping Connors–Auburn Arc and the surmised Baldwin Volcanic Arc (Murray et al., 1987; Scheibner and Veevers, 2000; Glen, 2005; Champion et al., 2009; Offler and Murray, 2011).

Latest Carboniferous–Early Permian oroclinal deformation of the SNEO most probably involved about 500 km northward displacement of the western Tamworth Belt versus the Lachlan Orogen (Section 4.5.6; Cawood and Leitch, 2006; Klotwijk, 2009; Pisarevsky et al., 2010; Cawood et al., 2011a, 2011b). Such movement is at the limit of paleomagnetic resolution and absence of correction would hardly affect pole path or paleolatitude representations. Structural grain and paleomagnetic data from the western Tamworth Belt do not indicate noticeable rotation with respect to cratonic Australia (Klotwijk, 1996a, 2009). The SNEO pole path and its implications for Carboniferous paleolatitudinal evolution (Fig. 5B; Supplementary Table 1) are taken therefore as broadly representing the Australian craton (but see Section 4.5.6 and Fig. 23A for corrected representation). There is good first-order agreement between the Carboniferous SNEO path (Fig. 5A) and the KG path for cratonic Australia (Klotwijk, 2009, fig. 2) indeed. The overall Australian Phanerozoic paleolatitudinal evolution (Fig. 6; Klotwijk, 1996a, figs. 17, 21) suggests an Early Devonian start for the Devonian–Early Carboniferous northward excursion. Such an early start would be contemporaneous with, and likely related to, start of closure of the Rheic Ocean between southwestern Laurasia and northwestern Gondwana (McKerrow et al., 2000; Winchester et al., 2006; Martínez Catalán et al., 2009; Nance et al., 2010; Zeh and Gerdes, 2010). The Tarim Block (Fig. 6), and possibly North China (Klotwijk, 1996a, fig. 21), underwent a Late Devonian–Early Carboniferous northward movement of comparable magnitude and timing as Australia. This has been interpreted as coherent northward movement of northeastern Gondwana/peri-Gondwana till collision with the CAOB (Klotwijk, 1996a, 2009, 2010), in contrast to prevalent global reconstructions (Fig. 1; Scotese, 2001; Golonka, 2002) featuring northward movements of peri-Gondwanan blocks separated from a more southern Gondwana. Early Devonian initiation of the northward excursion may have involved extension in peri-Gondwana, interpreted by Metcalfe (1996, 2011) as

Fig. 5. (A) Carboniferous pole path (blue) based on integrated data from the Rocky Creek (RC), Werrie (W) and Rouchel (R) Blocks of the western Tamworth Belt, Southern New England Orogen (Fig. 4B). Yellow = confidence ellipse around the pole position, or the envelope of ellipses, at the 95% confidence level. Legend: prim = primary magnetization; sec = secondary magnetization; N = Normal polarity; R = reverse polarity; M = mixed polarity; L2 = middle Carboniferous loop; L3 = latest Carboniferous–earliest Permian loop; SHRIMP U–Pb zircon ages in million years. The pole path is shown on a Gondwana reconstruction after Lawver and Scotese (1987), with Australia in its present position. (B) Carboniferous paleolatitudinal evolution for Armidale (151.67°E, 30.53°S, Southern New England Orogen) according to the pole path (A). Paleolatitudes are shown in order of their sequential position on the pole path, using relative pole declinations as a proxy for relative age control. SHRIMP-dated results provide indicative age. Legend: 1 = Waverley Fm.; 2 = Merlewood Fm.; 3 = Ayr Conglomerate Mb., Isismurra Fm.; 4 = Native Dog Mb. (primary magnetization, reverse polarity); 5 = Unnamed ignimbrites N and E of Albano; 6 = Unnamed ignimbrites near Albano; 7 = Native Dog Mb. (overprint, normal); 8 = High Valley andesite tuff; 9 = Martins Creek Ignimbrite Mb., Newtown Fm. (thermal treatment); 10 = Martins Creek Ignimbrite Mb. (liquid nitrogen, af, thermal); 11 = Curra Keith Tongue (overprint, normal); 12 = Yuendoo Rhyolite; 13 = Paterson Volcanics; 14 = Peri Rhyolite; 15 = tuffaceous beds above Boomi Rhyolite; 16 = Curra Keith Tongue (overprint, reverse); 17 = Koorngal Dacite; 18 = Rocky Creek Conglomerate; 19 = base Currabubula Fm., Kankool Basin; 20 = Lark Hill Fm., Rocky Creek; 21 = Lark Hill Fm., Tops Road; 22 = base Currabubula Fm., Werrie Basin; 23 = top Currabubula Fm., Werrie Basin; c = combined paleolatitude result for Lark Hill Fm., results 20 and 21; P/O = primary/overprint magnetization. Pole position data (1–23) detailed in Klootwijk (2002, table 4), Klootwijk (2003, table 4), Klootwijk (in preparation, table 4) and compiled in Supplementary Table 1.

start of the Paleotethys in contrast to the middle Carboniferous start advocated herein (Section 2.4).

2.3.3. Carboniferous paleolatitudinal evolution of the New Guinean cratonic promontory

The Australian craton extends northward up to the central New Guinean cordillera (Pigram and Panggabean, 1984; Klootwijk et al., 2003; Davies, 2012). The Carboniferous paleolatitudinal evolution of the SNEO (Fig. 5B) has been transferred to the central New Guinean cordillera

(Fig. 7D) in order to compare the paleolatitudinal evolution of the New Guinean cratonic promontory with the western CAOB. The New Guinean paleolatitudinal pattern is not corrected for about 500 km northward displacement of the SNEO versus the Australian craton during latest Carboniferous–Early Permian oroclinal deformation (see Section 4.5.6 and Fig. 23A for correction), nor for up to 150 km crustal shortening of central Australia during the Devonian–Carboniferous ASO (Section 4.4.2). Such corrections would result in more northern paleolatitudes than the about 20°N to 40°N range shown for the central New Guinean cratonic

Fig. 6. Early Paleozoic to Recent paleolatitudinal evolution of Australia (blue, Canberra: 149.1°E, 35.3°S reference) and of the wider Tarim Block (red) including the Junggar Basin (Urumchi: 87.6°E, 43.8°N reference) and the Kun Lun (Aksu: 80.3°E, 41.2°N reference). Note the resemblance of the Late Devonian–Early Carboniferous segments.

promontory for Early to middle Carboniferous (Fig. 7D). The New Guinean paleolatitude range overlaps with well-determined Devonian–Carboniferous (Kazakhstan Orocline) and Silurian–Devonian (Tuva terrane) paleolatitudes for the western CAOB (Fig. 7D; Supplementary Table 2), and also with age-equivalent paleolatitudes for central Mongolia (Kovalenko, 2010), the wider CAOB (Didenko et al., 1994) and wider Siberia (Cocks and Torsvik, 2007).

Paleomagnetism cannot define the longitudes of collision between the New Guinean promontory and the about east–west aligned western segment of the CAOB (Fig. 7A–C). Whereas global reconstructions (Scotese, 2001; Golonka, 2002) indicate collision between, previously separated, peri-Gondwanan fragments and the CAOB, the SNEO/Australian/Gondwanan and Tarim paleolatitude data (Figs. 5, 6) instead indicate collision of a coherent Gondwana–peri-Gondwana entity with the CAOB. Peri-Gondwanan reconstructions generally show Tarim located west of the New Guinean promontory (Scotese, 1987; Burrett et al., 1990; Metcalfe, 1996; Zhao et al., 1996; Metcalfe, 2011). Collision of the New Guinean promontory with the CAOB likely occurred eastward of Tarim's current location in the CAOB (Fig. 7B), probably with the Kazakhstan segment of the CAOB. Upon Late Carboniferous southward reversal of Gondwana/peri-Gondwana's movement, Tarim remained lodged within the CAOB (Fig. 7C), whereas the Cathaysian continents separated from eastern Gondwana at equatorial latitudes to rejoin the CAOB during Late Paleozoic–Early Mesozoic. This leaves Tarim, but not the Cathaysian continents, as an indicative constraint on longitude of collision between the New Guinean promontory and the CAOB. Faunal affinities between Early Paleozoic Kazakhstan and peri-Gondwana and also between Middle Paleozoic Kazakhstan (Chingiz Range) and eastern Australia (Eastern Lachlan Orogen) (Edgecombe, 1998; Popov et al., 2009; Klotwijk, 2010) can be interpreted in support of collision of the New Guinean promontory with the Kazakhstan segment of the CAOB, although observations are limited at present.

2.3.4. Central Asian Orogenic Belt concepts and paleolatitudinal control

Central Asia is dominated by a 1000 km wide, 7000 km long, orogenic complex that has grown outward from the Siberian craton

during mainly the Paleozoic. It is commonly referred to as Altaids Tectonic Collage (Şengör et al., 1993; Şengör and Natal'in, 1996), Ural–Mongol Fold Belt (Bazhenov et al., 2003; Van der Voo et al., 2006; Abrajvitch et al., 2007; Levashova et al., 2009) and Central Asian Orogenic Belt (CAOB) (Zonenshain et al., 1990; Didenko et al., 1994; Mossakovsky et al., 1994; Windley et al., 2007).

Hypotheses for evolution of the CAOB can be grouped twofold (see Xiao et al., 2010 for a more extensive overview). One group favors frontal subduction of oceans and back-arc oceans, collision of intra-oceanic arcs and accretion of continental blocks, largely of Gondwanan origin (Didenko et al., 1994), as the main process of continental growth (Zonenshain et al., 1990; Mossakovsky et al., 1994; Dobretsov et al., 1995). The other group stresses growth by duplicating subduction systems through large-scale strike-slip movements (Şengör et al., 1993; Şengör and Natal'in, 1996). Şengör and co-workers tied their evolutionary model to a broad paleolatitudinal framework (Fig. 7A–C). Recent paleomagnetic studies on evolution of the Kazakhstan Orocline by Van der Voo, Bazhenov and co-workers further detail this framework (Fig. 7D; Supplementary Table 2). Şengör's model is followed here for paleolatitudinal comparison with the Devonian and mainly Carboniferous northward excursion detailed by recent paleomagnetic data from the SNEO (Supplementary Table 1) that are considered representative for Australia as part of Gondwana (Sections 2.3.1, 2.3.2).

Şengör's model features evolution of a gigantic arc complex, the Kipchak and Tuva–Mongol Arcs, from Late Vendian to Cambrian–Ordovician initiation by separation from Baltica–Siberia to Early to middle Carboniferous cratonization with intense tectonization and telescoping of the arc complex. Cratonization is driven by convergence between Siberia and Baltica with accretion of Tarim by Early to middle Carboniferous and North China before Early Permian (Fig. 7A–C). Şengör's and many other studies on closure of the Paleasian Ocean, featuring cratonization of the CAOB – the Tianshan in particular – and accretion of Tarim, broadly agree on timing of major events. Main closure-related tectonism in the western CAOB occurred during Late Devonian–Early Carboniferous by Late Devonian–Early Carboniferous collision of Tarim with the Tianshan (Charvet et al., 2007; Sun et al.,

Fig. 7. (A–C) Evolution of the Kipchak–Tuva–Mongol arc complex, from its Vendian origin (A) to its Early Carboniferous (B) and Late Carboniferous (C) configurations as part of the Central Asian Orogenic Belt (CAOB), after Şengör and Natal'in (1996, fig. 21.28). (D) Carboniferous paleolatitude pattern for the central New Guinean cratonic promontory of Australia (blue with red squares, 141°E, 5°S reference), based on data from the Southern New England Orogen (Fig. 5; Supplementary Table 1). Paleolatitudes for the Kazakhstan Orocline (Devonian: 1–10, Carboniferous: 11–12) and Tuva Terrane (Silurian–Devonian: 13) of the western CAOB are shown for comparison (green squares, paleolatitudes as determined at individual locations, not converted to a common location and not strictly in stratigraphic order, data compiled in Supplementary Table 2). The 20°N to 40°N paleolatitude band is highlighted in orange. The CAOB data and the Australian data are not shown on a common time axis, the Silurian–Carboniferous age range for the CAOB data (yellow inset) is much wider than the Carboniferous range for the Australian data.

2008; Biske and Seltmann, 2010; Gao et al., 2011; Wang et al., 2011a), the Yili terrane of the Tianshan with the Junggar terrane (Gao et al., 1998; Zhou et al., 2008; Han et al., 2010; Zhang et al., 2012c), the Junggar terrane with the Altai–Mongolian terranes (Zhang et al., 2012b) and Kazakhstan with Siberia (Buslov, 2011; Chen et al., 2012b). There is however, no consensus on completion of Tarim–Tianshan collision, with timings ranging to Late Carboniferous (Wang et al., 2007; Yakubchuk, 2008; Gao et al., 2011; Han et al., 2011; Shu et al., 2011) and Permian (Xiao et al., 2009, 2010), possibly progressing diachronously (Han et al., 2011).

Şengör's model features a major structure in the western CAOB, the Kazakhstan Orocline, initiated during Devonian and Early Carboniferous by convergence of Siberia and Baltica and developed further during Late Paleozoic to Triassic strike-slip deformation (Allen et al., 1995; Natal'in and Şengör, 2005; Wang et al., 2007; Yakubchuk, 2008). In contrast to Gondwana-derived Tarim to the west and North China to the east, no Gondwana-derived indenter is identified by Şengör and co-workers in development of the Kazakhstan Orocline (Fig. 7B, C). Recent paleomagnetic studies (Supplementary Table 2) constrain oroclinal deformation between Middle Devonian and Middle Permian (Van der Voo, 2004; Abrajvitch et al., 2007; Levashova et al., 2009), with further strike-slip related rotations during Late Permian–earliest Mesozoic (Van der Voo et al., 2006). The northward excursion, defined above from SNEO paleomagnetic data (Sections 2.3.1, 2.3.2), adds Australia/northeastern Gondwana as a potential agent in deformation of the western CAOB and

permits speculation that indentation by the central New Guinean cratonic promontory (Section 2.3.3) contributed to initiating the Kazakhstan Orocline.

2.4. Paleasian–Rheic Ocean and Paleotethys concepts

Paleotethys is a loaded concept with a long history (Bernoulli and Lemoine, 1980; Jenkyns, 1980; Şengör, 1990; Tozer, 1990; Şengör and Atayman, 2009) and many different interpretations. Some do overlap with the Paleasian–Rheic Ocean concept. Several leading global reconstructions show the Paleotethys as a single wide ocean that remained in existence during Paleozoic between Laurentia–Baltica–Siberia–Kazakhstan and Gondwana (Heckel and Witzke, 1979; Scotese and McKerrrow, 1990; Yakubchuk, 2004). However, many studies on Paleozoic interaction between eastern Laurasia and eastern Gondwana distinguish at least two oceanic basins: an older, northern, *Paleasian Ocean* (another loaded term [Cocks and Torsvik, 2007] including, or relating to, the Turkestan, Junggar and Khanti–Mansi Oceans) between Siberia–Kazakhstan and Tarim–North China; and a younger, southern, *Paleotethys* between Tarim–North China and Gondwana (Zonenshain et al., 1990; Mossakovsky et al., 1994; Dobretsov, 2003; Buslov et al., 2004; Burtman, 2008). Many studies on Paleozoic interaction between western Laurasia and western Gondwana likewise distinguish an older, northern, *Rheic Ocean* (Murphy et al., 2006; Winchester et al., 2006; Bozkurt et al.,

Fig. 8. (A) Cenozoic impingement (green) of Gondwanan continental fragments — Adria, Arabia, India and Australia — into southern Eurasia, closing the Neotethys (after Klotwijk et al., 2003, fig. 1). (B) GPS velocity vectors for the Middle East and southern and eastern Asia in a Eurasian reference frame (after Becker and Faccenna, 2011, fig. 1A), shown in red for the Indian and Arabian indenters and in yellow for the Eurasian indentees.

Fig. 9. (A) Flow lines (magenta) across the Tibetan Plateau, based on kinematic interpretation of surface structures. Red arrows indicate direction and changes in speed of flow. Gray arrows indicate east component of velocity vector with respect to Eurasia. Blue arrows indicate velocity field in Tibetan reference frame (after Gloaguen and Ratschbacher, 2011, fig. 3 and Ratschbacher et al., 2011, fig. 18c). Legend: A-NBS = Assam-Namche Barwa syntaxis; GS = Gongha syntaxis; LMS = Longmen Shan; MFT = Main Frontal Thrust; QLS = Qinling Shan. (B) Dynamically maintained excess topography across the Longmen Shan and Sichuan Basin, compared with an average of topographic profiles from the northeastern and southeastern Tibetan Plateau toward the north and southwest of the Sichuan Basin (after Clark et al., 2005a, fig. 12). (C) Left: east–west section across the Longmen Shan and Sichuan Basin, idealized as a fault propagation fold. Pc = Precambrian crystalline basement; Pz = Paleozoic strata; ET = Early Triassic strata; LT-Jr = Late Triassic–Jurassic foredeep strata. Right: sketch relating upper crustal folding, faulting and high topography of the Longmen Shan to thickening of low-viscosity lower crust beneath eastern Tibet (after Burchfiel et al., 2008, fig. 7; see also Royden et al., 2008, fig. 6).

Fig. 10. (A) Centralian Superbasin (after Walter et al., 1995, fig. 1) and Larapintine Seaway (after Webby, 1978, fig. 6). (B) Central Australian Heat Flow Province (after McLaren et al., 2003, fig. 1b). (C) Comparison of surface heat flow for Australian Paleo-Mesoproterozoic High Heat Generating Granites (Oz Pr) with global averages (after Morgan, 1984, fig. 8 and McLaren et al., 2003, fig. 2). (D) Schematic crustal extension across the Larapintine Graben with pure shear in lower crust (LC) (after Lister et al., 1986, fig. 1a).

2008; Murphy and Nance, 2008; Nance and Linnemann, 2008) and a younger, southern, *Paleotethys* (von Raumer et al., 2002; Cocks and Torsvik, 2006; Stampfli and Kozur, 2006; Ribeiro et al., 2007). The Rheic and Paleasian Ocean concepts seem to have developed largely independently among researchers focussed on either eastern or western Laurasian or Laurussian interests. It is likely that the Rheic Ocean

originated, in Early Ordovician, as a western conjugate to a branch of the Paleasian Ocean.

With interpretations aplenty and at variance, application of the Paleotethys and Paleasian–Rheic Ocean concepts requires clarification. Usage herein relates to two major loops on the Australian Paleozoic pole path (Fig. 5A; Klotwijk, 1996a, 1998, 2009), L1 and L2, each

Fig. 11. (A) Topographic/bathymetric image of the Australian plate and surroundings (courtesy Geoscience Australia) with outline of the “compression box”. The outline is not defined to the south, the box may well have extended within Gondwana to central Antarctica (see Section 4.2.4). (B) Schematic overview of tectonic extrusion during the Alice Springs Orogeny. Red arrows indicate far-field compression from Australia–Laurasia convergence during Devonian–Carboniferous. Orange arrows indicate ductile flow of lower crust, mainly from the Larapintine Graben into the Thomson Orogen and internal Lachlan Orogen. Major ENE–WSW fault zones (red) guided about 200 km upper crustal eastward displacement of the Thomson Orogen and Northern New England Orogen. Yellow compartments (after Shaw et al., 1996, fig. 5) indicate heated, weak, crust of the Larapintine Graben and weak, originally oceanic, crust of the Tasman Orogenic System.

Fig. 12. (A) Rayleigh wave speed regionalization of Australia relative to a modified PREM model (after [Simons et al., 1999](#), fig. 14a). Black hatched line indicates Tasman Line. (B) Rayleigh wave velocity distribution across the Australian plate relative to the PREM model (after [Debayle and Kennett, 2000](#), plate 1a, reproduced by permission of American Geophysical Union and the authors). (C) Rayleigh wave speed anomalies of central and eastern Australia relative to the ak135 model (after [Fishwick et al., 2008](#), fig. 3a, reproduced by permission of American Geophysical Union). (D) Comparison of Tasman Lines (after [Direen and Crawford, 2003](#), fig. 1 and [Kennett et al., 2004](#), fig. 1b). S&V = [Scheibner and Veevers \(2000\)](#). (E) Variation of the neutral line in perturbation of shear-wave speed relative to the ak135 model as a function of depth, derived from surface wave tomography (after [Kennett et al., 2004](#), fig. 3a). (F) Horizontal gradient of surface wave velocity for central and eastern Australia (after [Fishwick et al., 2008](#), fig. 4a, *ibid.*). The four lines (solid, dotted, dashed and dash-dotted) indicate distinct gradient structures ([Fishwick, 2006](#); [Fishwick et al., 2008](#)), with the dashed line indicating changeover from continental-like to oceanic-like velocities well eastward from the “Tasman Line” and well inland from the eastern sea-board. (G) Surface wave velocity azimuthal anisotropy (after [Fishwick et al., 2008](#), fig. 5a, *ibid.*). Azimuths of solid blue lines indicate directions of fast polarization planes and their lengths indicate magnitudes of anisotropy. See (F) for gradient structures. (H) Rayleigh wave azimuthal anisotropy (after [Debayle and Kennett, 2000](#), plate 2a, reproduced by permission of American Geophysical Union and the authors). Azimuths of red line segments indicate fast directions, their lengths indicate strengths of anisotropy. (I) Present-day stress pattern of the Australian crust (after [Hillis and Reynolds, 2003](#), fig. 6).

reflecting major change in global plate movements. Opening of the Paleozoic–Rheic Ocean relates to the L1 loop with its apex at Cambrian–Ordovician boundary time, and to the ocean developing from earliest Ordovician between Gondwana and Laurussia/Laurasia (see caption to

[Fig. 1](#)). Closure of the Paleozoic–Rheic Ocean relates to onset and middle–Late Visian apex of the L2 loop. The northbound limb of the northward excursion defined by the L2 loop ([Fig. 5A, B](#)) reflects Devonian–Early Carboniferous convergence and collision of Australia/

Fig. 13. Compilation of Devonian–Carboniferous compressional (purple) and extensional (green) tectonic phases for regions within the “compression box” (Fig. 11A; Amadeus Basin, Ngalia Basin, Drummond Basin and Anakie Inlier, Bundock Basin, Wagga Block, Sydney Basin) and without (Bonaparte Basin, Canning Basin). Compiled from Shaw *et al.* (1994), Veevers (2000), and personal communications by R.J. Korsch and I. Laurie (Geoscience Australia, 2009).

Basin, Ngalia Basin, Drummond Basin and Anakie Inlier, Bundock Basin, Wagga Block, Compiled from [Shaw et al. \(1994\)](#), [Veevers \(2000\)](#), and personal communications by I

northeastern Gondwana with the western CAOB, closing the Paleasian–Rheic Ocean. Opening of the Paleotethys is reflected by the southbound limb of the northward excursion, indicating middle-late Visean reversal to southward movement of Australia/northeastern Gondwana, opening the Paleotethys in its wake.

The wider Paleotethyan opening process is reflected in Early- to-middle Carboniferous establishment of an extensional regime in northeastern Gondwana and southeastern Laurasia. This is well documented for the wider Greater India/Tibet/SE Asia region in late Tournaisian–early Visean initial rift-related tectonics in the Tethyan Himalaya ([Sciunnach and Garzanti, 2012](#)), Visean–early Namurian initial outpouring of the extensional, tholeiitic, Panjal continental flood basalt province in the Himalayan and southern Tibetan region ([Gaetani and Garzanti, 1991](#); [Pogue et al., 1992](#); [Garzanti and Sciunnach, 1997](#); [Chauvet et al., 2008](#); [Zhu et al., 2010](#)), (Middle–)Late Paleozoic rifting in the Qiangtang–Sanjiang, Qinling–Qilianshan–Kunlunshan and Gangdese–Himalayan orogenic systems ([Pan et al., 2012](#)) and Early Carboniferous opening of the Jinshajiang oceanic basin (oldest U–Pb ages 347–343 Ma; [Zi et al., 2012](#)) of the wider Tibetan Plateau. For the wider CAOB, opening of the Paleotethys is documented in late Early to early Late Carboniferous initiation of a post-collisional extensional regime in the Tianshan ([Xia et al., 2004](#); [Sun et al., 2008](#); [Zhou et al., 2008](#); [Biske and Seltmann, 2010](#); [Zhang et al., 2012c](#)), with documentation for northeastern Gondwana in latest Visean–Namurian opening of the Westralian superbasin ([Klootwijk, 1996a, 2009](#)).

Within the present study's region of interest in south central Asia, where did opening of the Paleotethys and break-up of eastern Pangea

occur? Potential definitions are by tracing the northernmost occurrence of cold water Gondwana fauna, outlining the northern margin of Gondwana, and/or by tracing changeover from Laurasian- to Gondwanan-sourced sediments, outlining where the Paleotethys closed. Remnants of the Paleotethys are traced to the boundary between the northern Tibetan Songpan-Ganzi and Qiangtang Blocks and, alternatively, to an internal structure of the Qiangtang Block, relying on either sedimentary provenance or faunal constraints. Sediment provenance data trace the relict Paleotethys in northern Tibet to the Songpan-Ganzi Basin/ (western) Jinsha Suture (Gehrels et al., 2011; Zhang et al., 2012e), probably continued in northeastern Tibet to the Jinsha-Garzê-Litang Suture based on continuity of the Yushu-Yidun volcanic arc (Reid et al., 2005; Yang et al., 2012b). Faunal data trace the northern margin of Gondwana in northern Tibet to the central Qiangtang metamorphic belt (Metcalf, 2006; Liu et al., 2011; Metcalfe, 2011) and in northeastern Tibet to the Changning-Menglian suture (Metcalf, 2002; Yang et al., 2011b; Pan et al., 2012), based on distinguishing Cathasian warm-water faunas from Gondwanan cold water faunas and glacio-marine deposits. Large-scale thrusting in the Qiangtang Block (Kapp et al., 2007; Wu et al., 2012) may complicate locating the relict Paleotethys.

3. Convergence across the Neotethys

3.1. Cenozoic India–Asia collision and Paleozoic Australia–Asia tectonics

Insights into Middle–Late Paleozoic Australia–Asia convergence and ASO-related intra-cratonic deformation processes may come from a

Fig. 14. Top left: ASO-related tectonic extrusion model (Fig. 11B) with green dots indicating locations referred to in Section 4.4.2 and figures A–G. Top right: Carboniferous paleolatitudinal evolution of the central New Guinean promontory of cratonic Australia (Fig. 7D). (A) Thrust and fold structures in central Australia (after Veevers, 1984, fig. 225 and Powell, 1984, fig. 3). (B) Schematic overview of mega-kinkband structures along the southeast coast of the Eastern Lachlan Orogen (after Powell, 1984, figs. 1, 2). (C) Outline of the Big Bend Megafold and Clarke River Fault Zone (after Bell, 1980, fig. 1 and Harrington, 1981, fig. 1). Bottom right: schematic Devonian–Carboniferous stratigraphy of the Bundock Basin (D), Texas beds (E), Northern New England Orogen (NNEO) (F), and Southern New England Orogen (SNEO) (G) (mainly after Jell and Playford, 1985; Jenkins et al., 1993; Scheibner and Veevers, 2000). Legend: (A) LT = Lander Trough, OB = Officer Basin; (C) BF = Burdekin Fault, C = Cairns, CRFZ = Clarke River Fault Zone, PF = Palmerville Fault, T = Townsville; (D) Bundock Basin: Bo = Boroston Fm., Bu = Bulgeri Fm., HC = Harry Creek Fm., TM = Teddy Mount Fm.; (E) TB = Texas beds; (F) NNEO: CAA = Connors and Auburn Arch, Y = Yarrol Belt; (G) SNEO: BA = Baldwin Volcanic Arc, TB = Tamworth Belt; Timescale: Fa = Famennian, Fr = Frasnian, N = Namurian, S = Stephanian, T = Tournaisian, V = Visean, W = Westphalian.

widely studied, ongoing, analog. Cenozoic convergence of Gondwanan fragments with southern Eurasia at the northern margin of the Neotethys, has much in common with Middle–Late Paleozoic convergence of Gondwana with southern Laurasia at the northern margin of

the Paleoasian–Rheic Ocean. Both processes relate to northward movement of Gondwana, or Gondwanan fragments, across an east–west ocean and collision with Asia. There is a notable difference, however. Cenozoic deformation studies focus on indentation and tectonic

Fig. 15. (A) ASO-related tectonic extrusion model (Fig. 11B). (B) Gondwana reconstruction (after Lawver and Scotese, 1987) with global extent of the Darling River Lineament–Cato Fracture Zone through to the Central Indian Tectonic Zone (based on Klootwijk et al., 1985 and Reeves et al., 2004).

Fig. 16. (A, B) Extent of marine transgressions based on ostracod data for early Early Devonian (A, *woschmidt* Zone and *pesavis* Zone) and late Late Devonian (B, *Early marginifera* Zone and *Late gigas/Late rhenana* Zone) (after Talent et al., 2000, endnote maps), superimposed on pattern of mega-shears guiding tectonic extrusion of the Thomson Orogen and NNEO (Fig. 11B). (C, D) Distribution of Early–Middle (C) and Late Paleozoic (D) I- and S-granitoids in eastern Australia (after Richards, 1980, fig. 14), see (A).

extrusion of the Eurasian *indentee* (Tapponnier, 1977; Tapponnier et al., 1982), but the present study on Middle–Late Paleozoic convergence focuses on intra-cratonic deformation of the Australian *indenter*, rather than the CAOB *indentee*.

India–Asia collision and evolution of the Tibetan Plateau are the more widely studied examples of ongoing continental indentation and tectonics at the northern margin of the Neotethys. Deformational observations from upper crustal GPS and deeper crustal geophysical data are presented and insights into driving forces, localized and/or continuum deformation models, and lower crustal ductile flow processes are discussed, as guidance for a novel and highly provocative interpretation of Middle–Late Paleozoic tectonic evolution of Australia.

3.2. Cenozoic indentation and tectonic extrusion of Eurasia

3.2.1. Gondwanan indenters

Continental indentation and tectonic extrusion have been recognized in various settings, following early insights into Arabia–Eurasia (McKenzie, 1972) and India–Asia (Molnar and Tapponnier, 1975) collisions. Four Gondwanan continental fragments, Adria, Arabia, India and Australia, crossed the Neotethys during Cenozoic, overrode their subducted oceanic lithosphere slabs, and continued their northward movement by indenting southern Eurasia (Fig. 8A; Klootwijk et al., 1992; Van der Voo et al., 1999; Replumaz et al., 2004; Hafkenscheid et al., 2006; Becker and Faccenna, 2011) with speeds increasing eastward along the Neotethyan collision zone (Reilinger and McClusky, 2011). Australia is in early stage indentation, with initial deformation of the Sunda Platform and formation of the Banda Arc (Rangin et al., 1999; Charlton, 2000; Pubellier et al., 2003; Richards et al., 2007; Hall, 2011). The other three Gondwanan fragments are in more advanced stages of indentation. Adria's indentation of southern Europe, in concert with an extensional regime controlled by Dinaric and Carpathian

subduction, has led to eastward tectonic extrusion of the Eastern Alps and Alpine–Carpathian–Pannonian Blocks (Tapponnier, 1977; Royden et al., 1983; Ratschbacher et al., 1991; Selverstone, 2005; Wöfler et al., 2011). Arabia's indentation of southwestern Asia has led to formation of the Anatolian highlands and Zagros ranges and westwards extrusion of the Anatolian–Aegean region, and to formation of the Hellenic (Aegean) and Cyprian (Tauric) Arcs in concert with slab rollback of the African (Nubian) plate (Şengör et al., 2005; Floyd et al., 2010; Howe and Bird, 2010; Le Pichon and Kreemer, 2010; Özeren and Holt, 2010).

India has indented southern Asia over about 3000 km (Patriat and Achache, 1984; Klootwijk et al., 1985; Dewey et al., 1989; Johnson, 2002; Replumaz et al., 2010) since initial India–Asia contact at equatorial latitudes near K–T boundary time (Klootwijk et al., 1992; Yin and Harrison, 2000; Chen et al., 2010; Cai et al., 2011; Zhang et al., 2012a), with far-field effects extending up to the Tianshan and the Altai mountains bordering the Siberian craton (Molnar and Tapponnier, 1975; Yin, 2010). Upon late Oligocene–early Miocene initiation of the Main Central Thrust and South Tibetan Detachment Zone (Powell and Conaghan, 1973; Burchfiel et al., 1992; Le Fort, 1996; Searle, 1996; DeCelles et al., 2001), contiguous with ending of major thrusting in the Lhasa and Qiangtang Blocks (Kapp et al., 2007; Wu et al., 2012), intra-continental subduction became the dominant deforming agent. This involved shallow subduction of Greater India below its own northern rim, the Tethyan Himalaya, and also the Lhasa Block and part or all of the Qiangtang Block (Nábělek et al., 2009; Chen et al., 2012a; Liang et al., 2012; Yue et al., 2012; Zhang et al., 2012d), or alternatively beneath nearly the entire N–S width of the Tibetan Plateau as indicated by paleomagnetic unwinding of the rotationally overthrust western Himalayan Arc (Klootwijk et al., 1985, fig. 1) and seismic probing of Indian lithosphere (Li et al., 2008, fig. 1) or its lithosphere–asthenosphere boundary (Zhao et al., 2010, fig. 1). The paleomagnetically and the seismically determined

Fig. 17. (A) Paleogeographic reconstruction of the Tasman Orogenic System during latest Devonian–earliest Carboniferous (after Powell, 1984, fig. 223). Legend: L–R–L = Longreach–Rome–Lismore lineament (after Evans and Roberts, 1980, fig. 1); waves = previous deformation; upward triangles = volcanic chain; playing hatching = flysch wedge. (B) Early Carboniferous reconstruction of eastern Gondwana (after Li and Powell, 2001, fig. 14). Legend: deep orange = emergent continent above sea level today; pale orange = emergent continent below sea level today; yellow = terrestrial sedimentary basin; blue = shallow marine; green = fold belt.

northern extent of Indian continental lithosphere agree well for western Tibet with shallow underthrusting beneath nearly the entire plateau up to the Jinsha Suture, but may diverge for central and eastern Tibet with some of the seismic studies indicating changeover from shallow to steeper underthrusting near the Bangong–Nujiang suture. Neogene large-scale continental indentation and subduction have led to increased uplift of the Tibetan Plateau and enhanced denudation around its margins (Wilson and Fowler, 2011), oroclinal deformation of the Himalayan Arc (Klotzwijk et al., 1985; Ratschbacher et al., 1994; Kapp and Guynn, 2004; Schill et al., 2004; Li and Yin, 2008) probably facilitated by southward extrusion through channel flow of ductile “middle” crust (Hodges, 2006; Searle et al., 2006; Law et al., 2011), eastward tectonic extrusion of Tibet by localized movement along large-scale shear zones (Tapponnier et al., 2001; Zhang and Klemperer, 2005; Hubbard and Shaw, 2009; Zhang et al., 2009; Robert et al., 2010) and/or distributed deformation driven by lower crustal channel flow (Clark and Royden, 2000; Burchfiel et al., 2008; Royden et al., 2008; Yao et al., 2010; Yang et al., 2012a), and injection of weak and thickened Tibetan lower crust beneath the Qaidam Basin akin to channel flow (Karplus et al., 2011). [Tibetan “middle” crust is referred to hereafter as “lower” crust, in order to acknowledge doubling of the Tibetan crust by Indian continental subduction and for consistency with description of Australian crustal structures (Section 4)].

3.2.2. Upper crustal movements

Ongoing indentation of southern Eurasia by Adria, Arabia and India and consequent tectonic extrusion are detailed from GPS monitoring of upper crustal movements, showing systematic large-scale deflection of movement vectors. Fig. 8B exemplifies deflection of Arabia’s northward movement into westward movement across the western Anatolian Plateau and southward movement of southwestern Anatolia and the southern Aegean region (Floyd et al., 2010; Le Pichon and Kreemer, 2010; Özeren and Holt, 2010; Becker and Faccenna, 2011; Nocquet, 2012). The figure also exemplifies deflection of India’s northward movement into eastward movement across the Tibetan Plateau, obstructed by the Sichuan Basin and Ordos Plateau, bifurcating

into eastward movement to the north of the Sichuan Basin along the Qinling Shan and southward movement into southeastern Asia (Zhang et al., 2004; Shen et al., 2005; Gan et al., 2007; Zhang and Gan, 2008; Becker and Faccenna, 2011).

3.2.3. Lower crustal/uppermost lithospheric mantle movements

Seismic tomography based on surface waves (Shapiro et al., 2004; Chen et al., 2009; Zhang et al., 2011), body waves (Lev et al., 2006; Zhao et al., 2010; Huang et al., 2011; Li et al., 2011; Soto et al., 2012), ambient noise and teleseismic surface waves (Yao et al., 2010), receiver-function inversion (Ozacar and Zandt, 2004; Sherrington et al., 2004; Sun et al., 2012) and deep seismic reflection (Ross et al., 2004), show consistent patterns of anisotropy in what is more likely Tibetan lower crust than uppermost lithospheric mantle (Li et al., 2011). Such anisotropy patterns are generally interpreted as “frozen” ductile flow lines (Park and Levin, 2002; Klemperer, 2006; Lloyd et al., 2011). The patterns transform progressively across the Tibetan Plateau, from north–south alignment near the Himalayan India–Asia interface toward east–west alignment within the central Tibetan Plateau (Sol et al., 2007; Zhang and Gan, 2008; Zhao et al., 2010; Huang et al., 2011), and from eastward in the eastern Tibetan Plateau (Li et al., 2011; Zhang et al., 2011) toward southeastward in the Yunnan–Sichuan region, based on Rayleigh wave phase velocity dispersion (Yi et al., 2010), ambient noise and teleseismic surface wave tomography (Yao et al., 2010), and inversion of focal mechanism data into drag forces acting on the base of the upper crust (Zhu and Shi, 2004). Such systematically deflected anisotropy/flow line patterns are interpreted as deflection of tectonically-extruded lower crustal ductile flow, away from northward indenting Greater India, eastward toward and around the Sichuan Basin into the Qinling Shan and southward into Southeast Asia. There is obvious coherence between the seismically defined patterns of lower crustal flow and the patterns of upper crustal movements determined from GPS observations (Sol et al., 2007; Soto et al., 2012). Anisotropy in the deeper lithospheric mantle, where determined separately, differs from the above directional pattern in the lower crust/uppermost lithospheric mantle and aligns grossly with the plate motion vector. Such lower

Fig. 18. (A) Australian crustal elements map (after Shaw et al., 1996, fig. 5) with outline of the Eromanga–Brisbane deep seismic profile (after Finlayson et al., 1990a, fig. 1). (B) Location of the deep seismic profile across the Adavale Basin and Drummond Basin–Anakie Inlier–Nebine Ridge (after Finlayson et al., 1990a, fig. 1). Legend (from west to east): E = Eromanga; BT = Barcoo Trough (now considered part of the Warrabin Trough, [Murray, 1994]); CF = Canaway Fault; CT = Canaway Trough; WT = Warrabin Trough; QT = Quilpie Trough; WT = Westgate Trough; R = Roma; BF = Burunga Fault; AI = Anakie Inlier; DB = Drummond Basin. The box refers to Fig. 20C. (C) Alternative schematic evolutionary models (white versus red arrows) for formation of the Quilpie Trough as a ramp syncline through differential movement between upper crust (UC) and lower crust (LC). White arrows indicate westward movement of upper crust versus lower crust as proposed by Leven et al. (1990, fig. 6a) and Finlayson et al. (1990a, fig. 15a). Red arrows indicate alternative relative movement pattern, eastward movement of lower crust with respect to upper crust, see Section 4.4.3.

crustal decoupling between movement patterns in the deeper lithospheric mantle and the upper crust further argues for ductile flow within the Tibetan lower crust (Li et al., 2011).

3.3. What drives evolution of the Tibetan Plateau?

Himalayan–Tibetan tectonic extrusion is often described as gravitational collapse of highlands that, through continental indentation and subduction, have grown to unsustainable heights (Shen et al., 2001; England and Molnar, 2005; Beaumont et al., 2006; Copley, 2008; Royden et al., 2008). Gravitational collapse of the Tibetan Plateau is regarded as facilitated by ductile flow in the lower crust and driven mainly by regional differences in gravitational potential energy (GPE) within the uppermost crust. Flow in the lower crust is guided by rigid continental blocks (Clark and Royden, 2000) and a rigid lower lid of subducted Indian lithosphere (Priestley et al., 2008). Absence of seismicity in central Tibetan lower crust (Sloan et al., 2011), widespread uplift of Tibetan rift flanks and lower crustal stress relaxation (Ryder et al., 2011), and flatness of the Tibetan Plateau are regarded indicative for low viscosity of the lower crust and efficacy of flow (Fielding et al., 1994; Masek et al., 1994; Bendick et al., 2008).

Tectonic contributions from boundary forces are acknowledged in many models (Liu and Yang, 2003; Becker and Faccenna, 2011), but often not as primary drivers. Co-alignment around the Tibetan Plateau between radial patterns of principal compressional strain and GPE gradients (Flesch et al., 2001; Liu and Yang, 2003; England and Molnar, 2005; Calais et al., 2006; Copley, 2008) supports the gravity-driven proposition. It is well possible that southward channel flow toward the eroding Himalayan Arc front (McKenzie et al., 2000; Beaumont et al., 2006; Grujic, 2006; Jamieson et al., 2006; Law et al., 2011) is governed directly by GPE gradient and consequent deviatoric stress patterns. However, it is harder to imagine that an essentially passive, upper crustal GPE-driven, deformation process would be solely responsible for progressive deflection of ductile flow patterns in the lower crust over up to 180° away from Indian and Arabian indenter vectors, or for buckling of the Tibetan Moho (Jin et al., 1994; Braitenberg et al., 2003; Shin et al., 2009). Instead, it is estimated from joint analysis of GPS, geologic and seismic anisotropy data (Flesch et al., 2005; Ghosh et al., 2006; Bendick and Flesch, 2007; Bendick et al., 2008; Wang et al., 2008a) that deformation of central Asia is driven for about half by GPE differences and about half by collisional boundary forces—India–Asia collision in the south and Pacific subduction in the east. Global mantle circulation modeling (Becker

Fig. 19. (A) Western segment of the Eromanga–Brisbane deep seismic profile (after Finlayson et al., 1990a, map 4), from Cooper Basin (CB, west) to Surat Basin (SB, east). Legend (from west to east): WT = Warrabin Trough; QT = Quilpie Trough; CT = Cooladdi Trough; WT = Westgate Trough; R = Roma. (B) Alternative interpretation of (A), with non-reflective upper crust and highly reflective lower crust highlighted. Red arrows indicate eastward ramping-up of lower crustal sections along crustal-penetrating, west-dipping, detachments and ramps. The Early Carboniferous Roma Granite and Late Devonian–Early Carboniferous granites of the Nebine Ridge may derive from partial melts that moved upward through the crust along roof (Nebine Ridge) and base detachments. A comparable origin is proposed for granites of unknown age above the Quilpie Ramp. Eastward ramping-up is thought driven by Alice Springs Orogeny-related tectonic extrusion, hydraulically transmitted through the lower crust by ductile flow as a Middle–Late Paleozoic analog to Neogene eastward extrusion of Tibet (Figs. 8B, 9A, 20A, B).

and Faccenna, 2011) indicates dominant importance of mantle drag for India's continuing northward movement during Cenozoic collision. In other Neotethyan collisions, GPE differences and boundary forces contribute about equally to the dynamics of the Anatolian–Aegean region (Özeren and Holt, 2010), and boundary forces contribute far more than GPE differences to extrusion of the Eastern Alps (Robl and Stüwe, 2005; Selverstone, 2005). Selverstone (2005) argues that gravitational collapse has been unduly emphasized as the main, if not sole, cause of late orogenic extension, and that the concept may have outlived its usefulness. Relative ease of determining GPE differences (Liu and Yang, 2003) may have led to undue emphasis.

Rather than merely reflecting a Tibetan GPE driver, systematic deflection of India's northward movement toward eastward and southward extrusion likely reflects hydraulic transmission of indenter forces through the ductile lower crust (Zhao and Morgan, 1987; Bird, 1991; Yang and Liu, 2009; Jamieson and Beaumont, 2011; Karplus et al., 2011). Spatial coincidence of decreasing eastwards extrusional speeds with widening of the movement channel in central Tibet and of increasing eastward and southward extrusional speeds with narrowing of the movement channel between the Namche Barwa and Gongha syntaxes of southeastern Tibet–Yunnan (Fig. 9A; Gloaguen and Ratschbacher, 2011; Ratschbacher et al., 2011) points to hydraulic, rather than GPE, movement forcing. Fig. 9A also shows substantial differences in eastward extrusional velocity along a north–south profile across the Tibetan Plateau, in spite of its remarkable flatness. This argues against a GPE-gradient driver, but would reconcile with a hydraulic driver and viscosity changes. Systematic re-alignment of lower crustal ductile flow along the regional pressure gradient would be aided by partial melt and/or aqueous fluids evident from seismic and magnetotelluric studies (Nelson et al., 1996; Alsdorf and Nelson, 1999; Unsworth et al., 2005; Klemperer, 2006; Yao et al., 2010), possibly localized to rift systems (Hetényi et al., 2011) or widely distributed across the Tibetan Plateau (Yang et al., 2012a).

3.4. Continuum versus localized deformation

GPS studies often portray tectonic extrusion of southern Eurasia as continuum deformation, but this is disputed for the India–Asia collision. Several analyses of Tibetan–Himalayan GPS data (Chen et al., 2004; Calais et al., 2006; Meade, 2007; Thatcher, 2007; Zhang and Wei, 2011) and wider kinematic studies (Replumaz and Tapponnier, 2003; Taylor et al., 2003; Liu and Bird, 2008) advocate a pattern of mixed deformation, with multiple semi-rigid blocks embedded in a matrix of localized deformation. Such upper crustal movement models verge toward concepts of localized shear along major fault zones, as advocated by Tapponnier and co-workers. More widespread and detailed GPS studies may define smaller and smaller block sizes verging to distributed deformation of the upper crust (Yuan et al., 2011). In line with this, many recent models for lower crustal movements verge toward ductile flow concepts, built on pioneering studies by Zhao and Morgan (1987), Wernicke (1990), Bird (1991) and advocated by Burchfiel, Royden and Clark. The latter's ductile flow concept has evolved into a model of channel flow of eastern Tibetan crust, “oozing” northeastward around the Sichuan Basin into the Qinling Shan and southeastward into Southeastern Asia (Section 3.5).

The two end-models, the localized shear fault model of Tapponnier and co-workers and the distributed ductile flow model of Burchfiel and co-workers, are not necessarily mutually exclusive. The two types of crustal behavior may verge into each other (Nocquet, 2012), reflecting lateral and depth variation in mechanical strength of the lithosphere (Wang et al., 2008a; Kirby and Harkins, 2010; Wang et al., 2011b). They may co-exist, representing different, upper and lower, crustal expressions of eastern Tibetan tectonic extrusion, with the upper crust carried and dragged by movement of the lower crust (Molnar, 1988; England and Molnar, 1991; Shen et al., 2001; Burchfiel, 2004; Zhang and Gan, 2008) and with the GPS vector pattern (Fig. 8B) reflecting upper crustal convolution of a lower crustal ductile flow process. Whatever deformation concept(s) for Neotethyan indentation and

Fig. 20. (A) Topographic map of the India–Asia collision zone with the red hashed region indicating lower crust added by eastward ductile flow from central Tibet (Clark et al., 2005a, fig. 15). (B) Topographic contour plot of the Tibetan Plateau and surrounding regions. SB = Sichuan Basin. Dark/light gray shading indicates cold strong/intermediate strength lithosphere. Tibetan lower (“middle”, see Section 3.2.1) crustal material escapes from the thickened plateau toward regions of weaker crust. Diverging red arrows are inferred directions of lower crustal flow (after Clark and Royden, 2000, fig. 5 and Royden et al., 2008, fig. 1). (C) Extensive, curved, band of folds across the western Drummond Basin (DB), draping onto and around the Nebine Ridge–Anakie Inlier (AI) (after Olgers, 1972, fig. 24). The fold band of middle Carboniferous (post-Ducabrook Fm., see Fig. 13) origin is eastward convex in the Bogantungan Embayment (BE) and eastward concave in the Scartwater Salient (SS), with tectonic grain changing from N–S toward E–W between the Anakie Inlier and Lolworth–Ravenswood Block (RA). Legend: BB = Bulgonunna Block; BoB = Bowen Basin; E/S B = Eromanga Basin/Surat Basin; GB = Galilee Basin; a = anticline; s = syncline. See Fig. 18B for location. Compare with Figs. 9C, 19B and 20A, B for analogy of the curved ramp-up and fold draping structure surfacing west of the Nebine Ridge–Anakie Inlier, with oozing of eastern Tibetan crust to the north and south around the Sichuan Basin and ramping-up in the Longmen Shan.

tectonic extrusion may prevail, the importance of lower crustal ductile flow and partial melt (Rosenberg and Handy, 2005; Kohlstedt and Holtzman, 2009; Vanderhaeghe, 2009; Whitney et al., 2009; Rutter et al., 2011) seems obvious from a wide spectrum of observations.

3.5. Tibetan lower crustal ductile flow processes

Studies of Tibetan lower crustal ductile flow concentrate on two types of channel flow, southward channel flow toward the Himalayan Arc and eastward channel flow toward and around the Sichuan Basin. Burchfiel, Royden, Clark and co-workers are developing an interesting, coherent, hypothesis for eastward expansion and extrusion of the Tibetan Plateau, based firmly on geological observations (Burchfiel, 2004; Clark et al., 2005b; Schoenbohm et al., 2006; Burchfiel et al., 2008) and geophysical modeling (Clark and Royden, 2000; Shen et al., 2001; Clark et al., 2005a; Royden et al., 2008). Their model advocates eastward outflow of ductile lower crust from the Tibetan Plateau, driven indirectly by GPE stored in the Tibetan Plateau (Clark and Royden, 2000; Burchfiel, 2004; Clark et al., 2005a), more directly by tectonic forces driving India's ongoing continental subduction and indentation of southern Asia, and aided by slab rollback in the western Pacific and Sunda regions (Sandiford et al., 2005; Royden et al., 2008; Yi et al., 2010). Lower crustal flow is guided by obstructing masses (Axen et al., 1998), Pamir in the west and Tarim in the north, into eastward flow. The flow is obstructed to the east, mainly by the Sichuan Basin but also the Ordos Plateau, diverging around the Longmen Shan and Min Shan, northward and eastward into the Qinling Shan and

southward into Yunnan (Fig. 9A, see also Fig. 20A, B). Eastward flow of Tibetan lower crust, facilitated by a network of forward-tunneling channel flows (Clark et al., 2005a; Harris, 2007; Burchfiel et al., 2008; Royden et al., 2008; Bai et al., 2010), results partly into eastward movement of upper crust it carries and drags (Zhu and Shi, 2004) and partly into thickening of lower crust (Fig. 9C) giving rise to eastward expanding uplift and thus eastward growth of the Tibetan Plateau.

The uplift aspect of Burchfiel and co-workers' lower crustal flow model is supported by most structural, geomorphic and GPS studies on the eastern Tibetan–Sichuan Basin region that show only limited Cenozoic and ongoing crustal shortening across the Longmen Shan with most deformation Mesozoic in age (Burchfiel, 2004; Royden et al., 2008; Wang et al., 2008b; Li et al., 2009; Zhang and Wei, 2011). Some studies however, point to more extensive Cenozoic shortening (Godard et al., 2009; Hubbard and Shaw, 2009; Zhang et al., 2009; Li et al., 2010a; Wilson and Fowler, 2011) or earlier, about K–T boundary time, crustal thickening (Wallis et al., 2003) and tend to question the lower crustal flow model. Further support for the model comes from body wave tomography and passive source seismic profiling across the Longmen Shan (Lei and Zhao, 2009; Li et al., 2009; Wang et al., 2010; Zhang et al., 2010) and from seismic tomography (Xu and Song, 2009; Li et al., 2010b; Wang et al., 2010; Li et al., 2012a; Zhang et al., 2012d) and magnetotelluric (Zhao et al., 2008; Teng et al., 2009) studies on the eastern margin of the Tibetan Plateau, showing a prominent low-velocity/low-resistivity zone in the middle-to-lower crust.

Excess height of the Higher Himalaya over the Tibetan Plateau, and of the Longmen Shan over neighboring regions with crustal flow oozing

Fig. 21. Note abrupt ending of Australia–Asia convergence/collision around 330 Ma (B), just before the start of the about 60 million year long Kiaman period of uniquely reverse polarity of the earth's magnetic field (A). Figures (C) to (I) compare lithospheric magnetic anomaly maps derived from satellite (C–H) and aeromagnetic data (I). (A) Duration of the Permo–Carboniferous Reverse Superchron, Kiaman, from about 327 Ma (Roberts et al., 2003) till about 265 Ma (Menning, 1995). M = mixed polarity, R = reverse polarity. (B) Schematic paleolatitudinal evolution for the central New Guinean promontory of the Australian craton during Carboniferous (see Figs. 5B, 7D, 14 top right). (C) Comprehensive model of the MAGSAT scalar residual field at 400 km altitude (after Purucker et al., 1997, fig. 1). (D) MAGSAT scalar anomaly field (contoured by Morgan, 2000, fig. 107 after vectorial data from Cain et al., 1989). (E) MAGSAT scalar anomaly field at 420 km altitude (after Ivers et al., 2000, fig. 11c). (F) ØRSTED scalar anomaly field at 780 km altitude (after Ivers et al., 2000, fig. 11a). (G) Comprehensive model CM4 of the ØRSTED and CHAMP scalar residual field at 400 km altitude (after Sabaka et al., 2004, fig. 14b). (H) Vertical component of the scalar residual field model FM6 with the lithospheric magnetic field estimated to degree 120 (after Maus et al., 2008, courtesy Peter Milligan, Geoscience Australia, personal communication 2008). (I) Long wavelength component (preliminary version: total magnetic intensity, upward continuation to 1 km, pass-band: 500–3000 km, reduced to the pole) of the magnetic anomaly map of Australia (Milligan and Franklin, 2004; Milligan et al., 2010, courtesy Peter Milligan, Geoscience Australia, personal communication 2012).

around the Sichuan Basin (Fig. 9B, see also Figs. 9A, 20A, B), can be regarded as a dynamic balance between forces exerted by gravity and by lower crustal flow (Clark et al., 2005a), hydraulically maintained by plunger action of India's continental underthrusting of southern Asia (Jin et al., 1996; Kola-Ojo and Meissner, 2001; Bendick and Flesch, 2007; Yang and Liu, 2009; Lee et al., 2011). Cenozoic deformation of the Longmen Shan was initiated in late Oligocene (Burchfiel et al., 2008; Royden et al., 2008; Roger et al., 2011) and its high topography developed in Late Cenozoic (Clark et al., 2005b; Meng et al., 2006; Godard et al., 2009; Kirby and Ouimet, 2011; Wilson and Fowler, 2011). The earlier event corresponds with late Oligocene–early Miocene initiation of intra-continental underthrusting of Greater India beneath its own northern rim along the Main Central Thrust, establishing

a hydraulic plunger effect in the sense of Zhao and Morgan (1987), Bird (1991), Jamieson et al. (2007), Jamieson and Beaumont (2011) and Karplus et al. (2011). This Tibetan lower crustal ductile flow scenario serves hereinafter as a conceptual guide for Middle–Late Paleozoic tectonic extrusion within Gondwanan Australia.

4. Alice Springs Orogeny and tectonic extrusion

4.1. Observational limitations

Findings of Australia-wide pervasive effects of ASO-related deformation keep emerging. The geodynamic driver for such large-scale Middle–Late Paleozoic tectonism has not been established beyond

Fig. 22. (A) Domains of high and low intensity magnetic crust in eastern Australia (after Wellman, 1998, fig. 7). (B) Magnetic domains and reworked magnetic bands in eastern Australia (after Wellman, 1995, fig. 2).

doubt, yet is crucial to unravel the Paleozoic evolution of Australia. New paleomagnetic insights on Middle–Late Paleozoic convergence of Australia with Asia and emerging insights on deformation of southern Asia from Cenozoic convergence of India with Asia are detailed above. What is this telling regarding cause and effect of the ASO? The challenge is to identify and analyze from geophysical and geological observations, potential evidence for upper crustal movements and lower crustal flow within the Australian craton and TOS during Middle–Late Paleozoic. This is a dire task in the absence of GPS and other kinematic observations on ongoing upper crustal movements that are so telling in analyzing Cenozoic extrusional deformation of southern Eurasia. The more so as Tibetan and other regional studies have shown that large-scale shear may not necessarily leave traceable upper crustal fault markers (Şengör et al., 1985; Burke and Şengör, 1986; Jones et al., 2006; Burchfiel et al., 2007; Raimondo et al., 2010). Absence of such kinematic observations in studies of Australian Middle–Late Paleozoic deformation requires increased reliance on the geophysical record of deeper crustal movements. What knowledge has accumulated that can be interpreted in the context of ASO-related tectonic extrusion?

There is no Australian precedent for the ASO Tectonic Extrusion Hypothesis (Section 4.3). Study of Mount Hay Ridge, central Arunta region, concluded deep crustal flow of channel or detachment flow origin during the Paleoproterozoic Stangways Orogeny (Bonamici et al., 2011). Several recent studies have argued for east-directed escape tectonics associated with the latest Neoproterozoic Petermann Orogeny in the central Australian, Arunta–Musgrave, region of intra-plate orogenesis (Aitken and Betts, 2009; Aitken et al., 2009a), and also for *southern Tibetan/Himalayan-type* channel flow involving lower crustal ductile flow and partial melt in the northern Musgrave Block (Raimondo et al., 2009, 2010). These studies seem soundly based on observational data, but do not identify a geodynamic driver, other than potential for a gravity-driven system. In contrast, the ASO Tectonic Extrusion Hypothesis argues for *eastern Tibetan-type* channel flow during the Middle–Late Paleozoic Variscan Orogeny, driven by convergence of Australia with Laurasia and facilitated by lower crustal ductile flow.

4.2. Tectonic extrusion — key elements

4.2.1. Formation of the Larapintine Graben

The structural elements map of Australia (Fig. 4A) features a prominent Central Australian mega-element, between the North Australian and South Australian–West Australian Cratons and extending from the Canning Basin east-southeast toward the TOS. It is made up of terranes accreted to the North Australian Craton during Paleoproterozoic (Betts and Giles, 2006; Payne et al., 2009) and is known also as the Amadeus Transverse Zone (Rutland, 1976) and the Central Australian Terranes (Myers et al., 1996). The central region of the Amadeus Transverse Zone acquired a substantial sediment cover as part of the Late Neoproterozoic Centralian Basin (Fig. 10A; Walter et al., 1995; Li and Powell, 2001), from initial basin formation around 800 Ma till the Petermann Ranges Orogeny (600–540 Ma), amounting up to 5.5 km in the Amadeus Basin (Wells et al., 1970; Flottmann et al., 2004) and up to 8 km in the southern Arunta Block (McLaren et al., 2009). The latter orogeny disrupted the Centralian Basin, accentuating previous partitioning (Camacho et al., 2002), into the Officer and Savory Basins to the south of the Musgrave Block, and a northern basins region that developed during Paleozoic into the Amadeus, Ngalia, and Georgina Basins. The Amadeus Basin and its wider extension across the Arunta Block acquired a further Cambrian sediment cover of up to 2.5 km (Wells et al., 1970).

A major structural change occurred during earliest Ordovician with initiation of the Canning Basin. Initial crustal extension coincided with initial opening of the Rheic Ocean (Murphy et al., 2006; Linnemann et al., 2007; Nance and Linnemann, 2008; Cocks and Fortey, 2009; Torsvik and Cocks, 2011), probably as a westward continuation of a branch of the Paleosian Ocean (Yakubchuk, 2008; Biske and Seltmann, 2010), between Gondwana and peri-Gondwana terranes (Stampfli et al., 2002; Keppie et al., 2003; Murphy et al., 2010; Nance et al., 2010; Yang et al., 2011a). It also coincided with the end phase of the Delamarian Orogeny (Foden et al., 2006), accretion of VanDieland (Cayley, 2011) and appearance of the Macquarie Arc (Glen et al.,

Fig. 23. (A) Comparison of the Carboniferous pole path for the Southern New England Orogen (SNEO, detailed in Fig. 5A and Supplementary Table 1) with this pole path corrected for 500 km northward movement of the western Tamworth Belt in forming the Texas and Coffs Harbour Oroclines. Interpreted primary/overprint (secondary) origin and polarity of corrected pole positions are indicated (legend as per [B]). L2, L3: see Fig. 5A. Pole path shown on Gondwana reconstruction (Lawver and Scotese, 1987) with Australia in its present position. (B) Compilation of primary (green) and overprint (red) poles for the Paleozoic of the "compression box" (Supplementary Table 3, New England Orogen poles excluded). Tentative pole path segments based on primary poles are indicated for Ordovician to Devonian and for Permian. The compilation highlights near absence of primary Carboniferous data, apart from one middle Carboniferous result for the Newcastle Range Volcanics (Anderson et al., 2003), and prevalence of overprints tentatively dated as Carboniferous from comparison with the SNEO pole path (A). Comparison of the overprint poles with the two primary pole path segments for the Paleozoic of the "compression box" indicates substantial presence of Permian overprints and less so Devonian overprints. Predominance of reverse polarity poles (primary and secondary) at or near the Permian pole path segment reflects the Permo-Carboniferous Reverse Superchron or "Kiaman". Short of an objective comparison of root mean square distances between "compression box" overprint poles and the original and the corrected SNEO pole paths, tentative visual comparison suggests a slightly better fit for the corrected path, based on: (i) Comparison of primary pole path segments for the "compression box" with the original and corrected SNEO pole paths seems equivocal across the older, Devonian–Carboniferous, interface, but indicates slightly better agreement for the corrected path across the younger Carboniferous–Permian interface; (ii) The Newcastle Range Volcanics pole, the only primary Carboniferous result for the "compression box", agrees marginally better with the corrected path; (iii) The same holds for a set of "compression box" overprints near the earliest Carboniferous part of the SNEO pole paths. Preference for the corrected SNEO pole path is interpreted as tentative support for the model of SNEO orocline formation detailed in Section 4.5.6 and Fig. 26. Pr = primary, S = secondary, M = mixed polarity, N = normal, R = reverse.

2007, 2012), and was probably driven by a major change in plate movements reflected in the L1 loop on the Australian Phanerozoic pole path (Klootwijk, 1998). Initiation of the Canning Basin during the Early Ordovician Sapphire Marsh Extensional Movement (Romine et al., 1994; Shaw et al., 1994) resulted in a trans-continental zone of extension, from the Canning Basin in the west, through the Amadeus Basin and Warburton Basin, to the Tasmanides in the east. This zone of extension facilitated western and eastern seaways (Veevers, 1976, 2000; Haines

and Wingate, 2007) that may have been connected as the Larapintine Seaway (Fig. 10A; Webby, 1978; Veevers, 1984; Nicoll et al., 1988; Walley and Cook, 1991; Li and Powell, 2001). The extension phase is identified in the deeper crust as a period of granulite-facies metamorphism, the Larapinta Event, dated around 480–460 Ma (Hand et al., 1999; Mawby et al., 1999; Buick et al., 2005; Maidment et al., 2006; Williams and Pulford, 2008). It is referred to here as formation of the Larapintine Graben. Development of the graben from Early Ordovician

to Early Devonian led to deposition of the Larapinta Group (Wells et al., 1970), adding a further 3.5 km to the sedimentary cover of basement in the Larapintine Graben.

4.2.2. Heat accumulation within the Larapintine Graben

Basement of the central Australian section of the Larapintine Graben forms part of the Central Australian Heat Flow Province (Fig. 10C; Cull and Conley, 1983; Morgan, 2000; Neumann et al., 2000; McLaren et al., 2003), with abundant Paleo- and Mesoproterozoic High Heat Generating Granites (HHGGs) (Fig. 10B; Neumann et al., 2000; Sandiford et al., 2002; Sandiford and McLaren, 2002; McLaren et al., 2005; Minty et al., 2010) and progressive enrichment of heat-producing elements in younger granites (McLaren et al., 2005; Sandiford and McLaren, 2006). Heat generated in upper crustal basement was conserved prior to and during the ASO (Cartwright et al., 1999; Hand et al., 1999; Mawby et al., 1999; Ballèvre et al., 2000; McLaren et al., 2003), blanketed by the Centralian Basin–Amadeus Basin sedimentary cover of up to 14 km including the Devonian Pertnjarra Group (Fig. 10A; Wells et al., 1970; Lindsay and Korsch, 1991; Walter et al., 1995; Haines et al., 2001; Flottmann et al., 2004). Such a substantial sedimentary blanket is supported by P–T modeling of Ordovician granulites from the Harts Range suggesting that its protolith volcano-sedimentary rift succession reached 30 km depth (Mawby et al., 1999; Scrimgeour and Raith, 2001; Maidment et al., 2004; Williams and Pulford, 2008). Meteoric fluids sourced from this sedimentary pile at mid crustal levels may have contributed to reduced crustal strength (Raimondo et al., 2012). Further heat may have been added to the lower crust by rising of uppermost lithospheric mantle into extending lower crust (Fig. 10D; Lister et al., 1986; Shaw, 1991; Shaw et al., 1991b; Morgan, 2000), altogether resulting in a heated, weakened crustal segment (McLaren and Sandiford, 2001; Flottmann et al., 2004; McLaren et al., 2005; Maidment et al., 2006), during later stages of the ASO in particular (McLaren et al., 2009). This hot and weak crust, with probably some partial melt, was primed for lower crustal ductile flow (Cull, 1989) upon impending Australia–Asia collision-related far-field stresses.

4.2.3. Substrate of the Tasman Orogenic System

Debate on substrate of the TOS has concentrated on the Lachlan Orogen, for limited exposure in the Thomson Orogen. The TOS is generally regarded as cratonized in Early–Middle Devonian by the Tabberabberan Orogeny (Veevers, 1984; Glen, 2005). Consensus has emerged that the Lachlan Orogen developed from an oceanic substrate (VandenBerg et al., 2000; Fergusson, 2003; Gray and Foster, 2004; Glen, 2005; Champion et al., 2009) and includes at least one continental fragment in central Victoria–Tasmania, variously identified as Victorian Microcontinent (Scheibner and Basden, 1998), Selwyn Block (VandenBerg et al., 2000; Cayley et al., 2002; Rossiter and Gray, 2008; Cayley et al., 2011), VanDieland (Champion et al., 2009; Cayley, 2011) and Tasmania–Selwyn block (Gibson et al., 2011).

Recent seismic evidence for the presence of continental fragments in basement of the Lachlan Orogen (Cayley et al., 2011) seems compelling, but extended discussions of isotopic evidence have come up with interpretations at variance. Petrological–geochemical arguments for a Precambrian continental substrate beneath parts of the Lachlan Orogen (Crawford et al., 1984; Chappell et al., 1988; Chappell and White, 1992; Scheibner and Basden, 1998; White et al., 1999) have been countered by crustal-sourced, complex, mixing models (Gray, 1984; Keay et al., 1997; Collins, 1998). Granites in the Eastern Lachlan Orogen have been related through zircon fingerprints to widely present Ordovician greywackes (Williams, 1992; Williams and Pulford, 2008) of a probably remote Gondwanan source (Williams et al., 2002; Squire et al., 2006). Chappell et al. (2000) and Williams and Pulford (2008) distilled from the prolonged debate that isotopic evidence refutes the presence of Precambrian crystalline shield beneath the Eastern Lachlan Orogen, but supports presence during Siluro-Devonian granite

formation of a thickened meta-sedimentary continental crustal protolith extending eastward up to the I–S line. The recent suggestion (Cayley et al., 2011) that I-type granites of the Stawell and northern Bendigo Zones relate to oceanic crust and S-type granites of the eastern Bendigo and Melbourne Zones relate to Selwyn Block continental crust may rekindle the isotope-centered debate on relations between granite types and composition of the deeper crust in the Lachlan Orogen. Debate on crustal origin of the Thomson Orogen (Draper, 2006) has evolved less because of limited exposure, but latest Neoproterozoic crust is well known in the Anakie Inlier and Charters Towers Province (Fergusson et al., 2009). Several extensive reviews (Murray, 1994; Glen, 2005) favor a flooring model for the Thomson Orogen of extended and thinned continental crust.

In spite of progressive Paleozoic cratonization of the eastern Australian TOS, its originally oceanic substrate would have remained hotter and weaker than the continental substrate of the North and South Australian Cratons (Royden, 1996; Goes et al., 2005). The TOS would thus have complemented the Larapintine Graben (Fig. 10A) as a heated and weakened zone at the time of the ASO. Combination of Middle–Late Paleozoic north–south compression with the weak and heated latitudinal zone of the Larapintine Graben, bounded to north and south by rigid cratonic blocks, and a weak and heated meridional TOS with an eastern “free oceanic boundary” where Paleopacific subduction rollback may have eased tectonic extrusion, arguably represent conditions comparable to Cenozoic tectonic extrusion in the Alpine–Himalayan belt (Section 3; Fig. 8A).

Purported presence of continental fragments within a TOS of original oceanic substrate, no doubt has contributed to the many diverging opinions about location and significance of the Tasman Line (Fig. 12D; Scheibner and Veevers, 2000; Direen and Crawford, 2003; Cayley et al., 2011; Withnall and Henderson, 2012; Korsch et al., 2012). It is generally taken as the eastern margin of the Australian craton upon Late Neoproterozoic–earliest Paleozoic transformation of Rodinia into Gondwana (Dalziel, 1997; Veevers, 2000; Li and Powell, 2001; Cawood, 2005). The spread of Tasman Lines does not seem to reflect first-order changes in depth to the base of the lithosphere (Kennett and Blewett, 2012). Scheibner and co-workers (Scheibner, 1992; Scheibner and Basden, 1998; Scheibner and Veevers, 2000) rationalized the spread of Tasman Lines and blurred distinction between continental and oceanic velocities in the internal TOS with the “Tasman Toeline” concept, an apparent outgrowth of the lower and upper plate concept of passive margin development (Lister et al., 1991). Alternatively, as argued hereafter (Sections 4.3, 4.4), continental-like velocities for basement of the internal TOS may reflect tectonic extrusion of ductile lower crust from the Larapintine Graben.

4.2.4. Compressional deformation confined within “compression box”

Convergence of Australia/northeastern Gondwana with the CAOB may have started in Early Devonian (Section 2.3; Fig. 6), with final collision in Early Carboniferous. The Australian cratonic promontory in central New Guinea probably spearheaded collision. Tangible convergence-related compressional deformation in Australia seems confined to a “compression box”, extending southward from the central New Guinean cratonic promontory (Fig. 11A). Its western boundary extends southward between the Bintuni Basin of western New Guinea and the Arafura, Bonaparte and Ord Basins, through the Halls Creek Fault Zone—the East Kimberley Geosuture of Rod (1966)—into the enigmatic Lasseter Shear Zone (Forman and Shaw, 1973; Wellman, 1976; Braun et al., 1991; Haines et al., 2001; Aitken, 2010). It is comparable to the G12 gravity lineament (Elliot, 1994). Its eastern boundary extends from the Aure Trough of eastern New Guinea—the Aure Geosuture of Rod (1966)—through the offshore Queensland Basin and the onshore Bowen–Gunnedah–Sydney basin system into the Tasman Sea (see also Veevers, 2000, fig. 318). It traces the Early Permian East Australian Rift System (EARS; Korsch et al., 2009b) and an eastern Gondwanan shear system (Visser and Praekelt, 1998; Cawood and Buchan, 2007; Klotwijk, 2009). The “compression box” is open to the south. Within Gondwana it may well have

Fig. 25. Late Devonian paleogeography of eastern Gondwana (after Li and Powell, 2001, fig. 13b), showing sediment transport from the Alice Springs Orogen toward the Canining Basin in the west and the Drummond and Lambian Basins in the east. Legend: orange = emergent continent above sea level today; pale orange = emergent continent below sea level today; yellow = terrestrial sedimentary basin; blue = shallow marine; A = Amadeus Basin; D = Drummond Basin; HC = Halls Creek Mobile Zone; L = Lambian Basin; N = Ngalia Basin; RH = Ross High.

extended into central Antarctica where mid-Carboniferous uplift of the Gamburtsev Subglacial Mountains has been attributed to Gondwana–Laurasia interaction (Veevers, 1994; Veevers et al., 1994a). Outside this “compression box”, there is much weaker Middle–Late Paleozoic compressional tectonics in the NEO (Korsch, 2004; Glen, 2005; Champion et al., 2009) and extensional tectonics in western Australia. The heated, weak, Larapintine Graben traverses the “compression box” from west to east, toward the hot and weak oceanic substrate of the TOS with Zealandia extension (Sdrolas et al., 2003; Mortimer, 2006) and the “free oceanic boundary” of the Paleopacific.

4.3. Tectonic extrusion model

4.3.1. Applicability

The ASO-related tectonic extrusion model is schematized in Fig. 11B. Detailed paleomagnetic control on the northward excursion of Australia/northeastern Gondwana is restricted to Carboniferous (Figs. 5B, 6, 7D). The model applies therefore essentially to Early Carboniferous tectonic phases of the ASO—the late Tournaisian–early Viséan Brewer–Kerridy Movement and the late Viséan–early Namurian Meda–Alice Springs–Mount Eclipse Movement. Less detailed Australian

Paleozoic paleomagnetic control (Fig. 6) suggests an Early Devonian start for the northward excursion, perhaps coinciding with start of closure of the Paleasian–Rheic Ocean (McKerrow et al., 2000; Martínez Catalán et al., 2009; Nance et al., 2010; Gutiérrez-Alonso et al., 2011). Application of the model may therefore be extended to Devonian tectonic phases of the ASO—the late Famennian Van Emmerick–Pre-Brewer–Kerridy phase, the Frasnian Henbury phase and the Emsian–early Eifelian Pertnjara phase (Fig. 13). This makes the model applicable across the conventional duration of the ASO (Wells et al., 1970; Bradshaw and Evans, 1988; Shaw and Black, 1991; Braun and Shaw, 2001; McLaren et al., 2009), from about 400 Ma (Pertnjara Movement) to about 330–320 Ma (Mount Eclipse Movement).

Paleomagnetic control on latitudinal movements of Australia/northeastern Gondwana prior to Devonian is more fragmentary. The paleomagnetic record has not yet been interpreted in terms of a possible geodynamic driver for the Late Ordovician–Early Silurian compressional phase of the extended ASO (Section 2.2.2) that led to amphibolite-facies metamorphism in the Harts Range (450–440 Ma: Hand et al., 1999; Mawby et al., 1999; Haines et al., 2001; Scrimgeour and Raith, 2001; Maidment et al., 2007), to reactivation of the Redbank Thrust system (450–420 Ma: Biermeier et al., 2003) and to the Rodingan Movement in the Amadeus Basin (430–440 Ma: Wells et al., 1970; Lindsay and Korsch, 1991; Shaw et al., 1991a; Walley and Cook, 1991). For want of a geodynamic driver, application of the tectonic extrusion model is not extrapolated herein across the extended ASO back to 450 Ma (Shaw, 1991; Dunlap et al., 1995; McLaren et al., 2009). However, correlation between tectonic events across regions affected by tectonic extrusion—Alice Springs Orogen, Warburton and Adavale Basins—has been pointed out by McKillop et al. (2007), for the mid-Carboniferous and Early–Late Devonian events of the conventional ASO range and also for the Late Ordovician event of the extended ASO range.

4.3.2. Outline of the ASO-related tectonic extrusion model

Devonian–Carboniferous northward convergence of Australia with Laurasia induced north–south to northeast–southwest far-field compression within the “compression box” (Figs. 3, 5, 11A) and, together with GPE deviatoric stresses, led to ductile flow in the weakened and heated lower crust of the Larapintine Graben. Lower crustal ductile flow was hydraulically driven, with re-alignment of flow along the local pressure gradient, progressively deflected eastward toward the weak, heated, substrate of the TOS and the “free oceanic boundary” of the Paleopacific. Lower crustal flow carried/dragged its upper crust eastward, bounded to north and south by major crustal-penetrating mega-shears guiding eastward extrusion of the Thomson Orogen and NNEO (Fig. 11B), with subduction rollback potentially easing extrusion. The northern mega-shear extends from shear systems in the eastern Arunta Block, along the Diamantina River Lineament and the Clarke River Fault Zone, into the offshore Townsville Trough. The main southern mega-shear extends along the Darling River Lineament and nearby Cobar–Inglewood Lineament, offshore into the Cato Fracture Zone. Movement is sinistral for the northern mega-shear and dextral for the southern mega-shear. A subsidiary southern mega-shear extending along the Lake Blanche Fault Zone and Olepoloko Fault Zone into the Lachlan Transverse Zone shows sinistral and dextral movements.

Seismic and magnetic images provide support for southward flow of lower crust through the internal TOS, mainly the Western and Central Lachlan Orogens (Sections 4.4.1, 4.5.1). Eastward and southward flow of ductile lower crust effectuated differently. Eastward flow toward

Fig. 24. (A) Filtered (pass-band <500 km) Bouguer gravity anomaly map (after Bacchin et al., 2008, courtesy Peter Milligan, Geoscience Australia, personal communication 2009). (B) Average differential horizontal stress field ($\sigma_{Hmax} - \sigma_{Hmin}$, after Reynolds et al., 2002, fig. 6, modified by permission of American Geophysical Union), with overlay of mega-shears for the ASO-related tectonic extrusion model (Fig. 11B). (C) Estimated crustal temperature at 5 km depth (courtesy Prame Chopra, Earthinsight, personal communication 2007, see also Chopra and Holgate, 2005, fig. 6). (D) Sediment thickness map (after OZ SEEBASE™, 2005, pp 134). (E) Electrically conductive structures (after Chamalaun and Barton, 1993, fig. 2 and Lilley et al., 2003, fig. 1), draped over an image of Phanerozoic basins (after OZ SEEBASE™, 2005, p 130, depth to basement map). (F) Fourth edition of the magnetic anomaly map of Australia (Milligan and Franklin, 2004; Milligan et al., 2004, fig. 5), showing gradient-enhanced residuals of total magnetic intensity (courtesy Peter Milligan, Geoscience Australia, personal communication 2008).

Fig. 26. Evolutionary model for the oroclines of the Southern New England Orogen (SNEO, see Section 4.5.6) in Gondwana context (Lawver and Scotese, 1987), highlighted through four relocations according to selected paleomagnetic data from the SNEO pole path (Fig. 5A; Supplementary Table 1), detailed mainly in Klootwijk (2003, table 4). (A) Waverley Formation (Klootwijk, in preparation: pole longitude = 335.4°E, pole latitude = 67.3°S, $K_p = 80.9$, $E_p = 8.6^\circ$). (B) Currabubula Formation, base. (C) Currabubula Formation, top. (D) Boggabri Volcanics.

the “free oceanic boundary” of the Paleopacific resulted in eastward displacement of lower and upper crust and thickening of lower crust. In contrast, southward flow, far away from the weak Paleopacific boundary and nearing the Selwyn Block barrier, would have resulted in mainly thickening of lower crust. Eastward flow of lower crustal material through the northern Thomson Orogen may have entrained some northward flow between the Mt Isa Block and the Georgetown Block in “tension gash”-like reactivation of weakened crust along the Carpentaria Conductor (Sections 4.4.1, 4.5.5). Westward flow of ductile lower crust from the Larapintine Graben into the Canning Basin probably was of lesser magnitude for absence of a nearby “free oceanic boundary”, with the Canning Basin confined within northern Gondwana and peri-Gondwana.

4.4. Supporting observations

4.4.1. Lower crustal evidence

Seismic tomography of the Australian region from surface wave (Fig. 12A–C, E, F; van der Hilst et al., 1998; Simons et al., 2002; Debayle and Kennett, 2003; Kennett et al., 2004; Fishwick et al., 2008), body wave (Graeber et al., 2002; Kennett, 2003; Rawlinson et al., 2010) and full waveform (Fichtner et al., 2010) data, consistently show higher velocities for the upper lithospheric mantle/lower crust of the internal than the external TOS. Although temperature variations may be involved, these higher velocities have been interpreted as, in part thinned, continental rather than oceanic crust (Glen and Walshe, 1999; Glen, 2005). This may be somewhat surprising given current consensus for an original oceanic substrate for the TOS, but exemplifies the prolonged debate about the presence of continental basement and recent seismic support for the Selwyn Block (Section 4.2.3). Regions with continental-like velocities cover the internal Thomson and Lachlan Orogens.

Seismic tomography has not yet shown velocity distributions specifically for lower crustal levels, with the uppermost maps centered

at uppermost lithospheric mantle depths of 50 km (Fishwick and Rawlinson, 2012), 75 km (Fishwick et al., 2005; Fishwick and Reading, 2008; Fichtner et al., 2010; Fishwick and Rawlinson, 2012), 80 km (van der Hilst et al., 1998; Debayle and Kennett, 2000; Simons et al., 2002; Fichtner et al., 2009), or 100 km (Debayle and Kennett, 2003; Yoshizawa and Kennett, 2004; Fishwick et al., 2008). The maps show velocity distributions relative to a global velocity reference model (PREM: Dziewonski and Anderson, 1981; modified PREM: Zielhuis and Nolet, 1994; ak135: Kennett et al., 1995), relative to the average velocity in the region (Fishwick, 2006), or as relative lateral variations (Fichtner et al., 2010). Structural resolution of surface wave tomography currently is no better than 200 km horizontal and 25 km vertical (Kennett et al., 2004), or 3° lateral and 40 km vertical (Fichtner et al., 2010), with optimum horizontal resolution approximately 100–200 km (Fishwick and Rawlinson, 2012). Comparative velocity maps are therefore not ideal to identify regions with continental or oceanic velocities.

Regions bounded by pronounced velocity changes may be better identified from horizontal velocity gradient maps (Fishwick, 2006; Fishwick et al., 2008; Fishwick and Rawlinson, 2012) that do not rely on a reference model and can highlight features not evident otherwise. Notably, such maps do demonstrate a prominent gradient in surface wave velocity between external and internal regions of the TOS (Fig. 12F, dashed line) that shifts substantially westward with depth (Fishwick and Rawlinson, 2012) and do support earlier tomography observations of continental-like velocities for the internal TOS (Fig. 12A–C, E). Full waveform tomography supports the above findings from surface and/or body wave analysis for potential presence of craton-like basement (Fichtner et al., 2009, 2010). Presence of continental-like velocities in uppermost lithospheric mantle/lower crust of the internal TOS also has been acknowledged in geological reviews (Glen and Walshe, 1999; Glen, 2005) and is supported by Os isotopic data for xenoliths from the Newer Volcanics in southeast

Fig. 27. (A) Overview map of the Tasman Orogenic System with major fault traces and aeromagnetic trends (after Gray and Foster, 2004, fig. 1). (B) Gradient-enhanced aeromagnetic anomaly map of eastern Australia (after Gray and Foster, 2004, fig. 3). The magnetic image is from Milligan and co-workers (Milligan P.R., Tarlowski, C., Mackey, T., 1999. Magnetic anomaly grid of Australia. [Digital Dataset]. Geoscience Australia, Canberra), with superimposed major fault traces (bold lines), structural trends, form lines, and aeromagnetic trend lines.

Australia indicating Early Proterozoic continental lithospheric mantle, and most likely crust, extending some 400 km eastward of the Tasman Line (Handler and Bennett, 2001).

How could such continental-like velocities have materialized? The maps nominally represent the uppermost lithospheric mantle, but may reflect some smearing of crustal velocities (Fishwick et al., 2005; Fishwick and Reading, 2008). The “Jelly Sandwich/Christmas Tree” strength profile of the lithosphere (Afonso and Ranalli, 2004; Burov and Watts, 2006; Bürgmann and Dresen, 2008; Chen et al., 2012a) points to the lower crust as the more likely channel for ductile flow and more likely origin for the continental-like velocities. A contrasting weak continental lithospheric mantle model, “Crème Brûlée”, has been argued for on seismogenic and gravity grounds (Maggi et al., 2000; Jackson, 2002), but its validity has been questioned in favor of the “Jelly Sandwich” model (Burov and Watts, 2006; Burov, 2010). This supports the lower crust, over lower crust to uppermost lithospheric mantle, as the more likely level of channel flow [in the present context of the ASO Tectonic Extrusion Hypothesis, “channel” is regarded as a crustal level with minimum viscosity, localizing horizontal flow (Gloaguen and Ratschbacher, 2011)]. Velocity distribution maps indicating continental-like velocities in the internal TOS, centered nominally on the uppermost lithospheric mantle, are interpreted therefore herein primarily in terms of eastward and southward extrusion of lower crustal material (Fig. 11B).

Surface wave azimuthal anisotropy studies show for central and eastern Australia eastward fanning out patterns at depths of 80 km (Simons et al., 2002; Simons and van der Hilst, 2003), or 100 km (Fig. 12G, H; Debayle and Kennett, 2003; Debayle et al., 2005; Fishwick et al., 2008). Body wave studies (Clitheroe and van der Hilst, 1998; Simons et al., 2002; Heintz and Kennett, 2005; Fishwick et al., 2008; Kennett and Blewett, 2012) show comparable anisotropy patterns. There is also notable correspondence across eastern Australia, the Thomson Orogen and internal Lachlan Orogen in particular, between the region of eastward fanning out of azimuthal anisotropy (Fig. 12H) and the region of lower crustal extrusion surmised from both seismic tomography (Fig. 12A–G) and deeper crustal magnetic data (Section 4.5.1; Fig. 21). Seismic anisotropy in the upper lithosphere is generally interpreted to reflect “frozen” ductile flow (Park and Levin, 2002; Simons and van der Hilst, 2003; Debayle et al., 2005; Fouch and Rondenay, 2006; Pawlak et al., 2012; see also Section 3.2.3). The eastward fanning out patterns of azimuthal anisotropy are interpreted therefore to reflect ductile flow in the lower crust. There is a notable precedent to such an interpretation in obvious support of the ASO Tectonic Extrusion Hypothesis. Debayle and Kennett (2000) pioneered attributing east–west alignment of anisotropy patterns in central Australia to north–south compression-driven “lateral extrusion” during the ASO indeed, citing personal communication to Debayle and Kennett

from J. Braun 1999. Resemblance between these fanning out anisotropy patterns and the present-day stress pattern of the Australian crust (Fig. 12I) may reflect the repetitive spearheading role by the cratonic promontory of central New Guinea (Fig. 11A) during Middle–Late Paleozoic Australia–Asia convergence and present-day Australia–Sunda Platform–Pacific convergence.

P–T modeling of mainly Paleozoic xenoliths from southeastern Australia (Griffin et al., 1984; O'Reilly and Griffin, 1985; O'Reilly et al., 1998) points to past temperature gradients in the lower crust substantially in excess of present-day surface heat flow measurements. This suggests an advective heat source in the lower crust at the time of xenolith extrusion (Griffin and O'Reilly, 1987; Lister and Etheridge, 1989; Cull et al., 1991; Gray and Cull, 1992). Advective heat sources are commonly interpreted as basaltic magma underplated to the lower crust (Drummond and Collins, 1986) and this may relate to sub-horizontal layering interpreted from high seismic reflectivity (Finlayson et al., 1993). Other suggested advective heat sources are shear-heating along middle and lower crustal detachments and rising of upper lithospheric mantle into extending crust. Alternatively, the advective heat source can be interpreted as heated, and perhaps partially melted, lower crust transported by ductile flow (Section 3.5). Sm–Nd model ages for xenoliths from Bullenmerri and Gnotuk in western Victoria suggest a lower crustal/uppermost lithospheric mantle origin around 300–500 Ma (Griffin et al., 1984, 1988). This age range is supported by U–Pb zircon ages for xenoliths from Calcutturoo, Flinders Ranges, indicating a youngest tectonic event in the lower crust/uppermost mantle at 330 Ma (Chen et al., 1994).

Griffin and O'Reilly (1987; Griffin et al., 1987) also studied an extensive suite of lower crustal xenoliths from central-eastern Queensland, showing P–T data in accordance with the high paleo-geotherm established for southeastern Australia. No age data are available for the xenoliths, although a Late Paleozoic age is suggested. Rudnick and co-workers (Rudnick and Williams, 1987; Rudnick, 1990; Rudnick and Goldstein, 1990) studied lower crustal xenoliths from the McBride volcanic province of the Georgetown Block in North Queensland. These show Nd model ages indicating possible derivation from Lower Proterozoic crust, equilibration at mid-to lower crustal depths at temperatures too high for normal geothermal gradients (Kay and Kay, 1983) and U–Pb zircon crystallization ages between 300 and 400 Ma, and point altogether to an intense Late Paleozoic lower crustal heating event (Rudnick, 1988). This age range is supported by Rhenium-depletion model ages of 350 Ma and 440 Ma for metasomatic events in lithospheric mantle-derived xenoliths from the Atherton Province and the Chudleigh Province respectively (Murgulov et al., 2011), studied previously by Rudnick. The xenolith data apparently record an advective heat event in lower crust of the wider TOS during the ASO (450/400–330 Ma). This may reflect transport of heated continental-like material through the lower crust of the internal TOS, in agreement with seismic tomography (Figs. 11B, 12) and magnetic anomaly (Section 4.5.1; Figs. 21, 22) interpretations.

4.4.2. Upper crustal evidence

Powell (1984) noted correspondence in time between Early to middle Carboniferous crustal shortening throughout the TOS and throughout central Australia, linking the shortenings to a north–south compressive event of mid-Carboniferous age but unknown cause. He estimated ASO-related crustal shortening in central Australia at 100 to 150 km over the 360–320 Ma period (Fig. 14A; Veevers, 1984) and crustal shortening across the 3000 km north–south extent of the TOS at 150 km over the 350–320 Ma period (Fig. 14B, C; Powell et al., 1985).

The outlines of the northern and southern megashears that bound ASO-related eastward extrusion of the Thomson Orogen and NNEO (Figs. 11B, 14 top left, C) derive from structural observations, but

seem reflected also in ambient noise Rayleigh wave group velocity tomography in the 20 second period range indicative for mid-crustal structures (Saygin and Kennett, 2011, figs. 5e, 9e,f). Sense of movement along the mega-shears is well established, sinistral along the northern mega-shear and dextral along the southern.

The northern mega-shear extends from the eastern Arunta Block, and possibly the western Arunta Block (McLaren et al., 2009), along the Diamantina River Lineament and Clarke River Fault Zone (CRFZ), into the offshore Townsville Trough (Struckmeyer and Symonds, 1997). It corresponds with the G1 gravity lineament (O'Driscoll, 1983). Sinistral shear of ASO vintage is well established throughout the Arunta Block (Collins and Teyssier, 1989), the Reynolds Range (Cartwright et al., 1999; Hand and Buick, 2001), Harts Range (Buick et al., 1998; Scrimgeour and Raith, 2001) and Strangways Range (Ballèvre et al., 2000; Haines et al., 2001) of the eastern Arunta Block in particular. Sinistral shear is also well established throughout the northern Amadeus Basin (Veevers, 1984; Bradshaw and Evans, 1988) and along the CRFZ (Arnold and Fawcckner, 1980; Henderson, 1980, 1987; Withnall and Lang, 1990; Bain and Draper, 1997; Withnall and Henderson, 2012). Australia-wide review studies also indicate sinistral movement near the northern mega-shear (Li and Powell, 2001; OZ SEEBASE™, 2005) and sinistral movement along the CRFZ is implied in Bell's (1980) 'Big Bend' model and Harrington's (1981) alternative triple junction model (Fig. 14C). Sinistral movements along the CRFZ probably started in Early Devonian (Withnall and Lang, 1990), with unspecified Late Devonian (Withnall and Henderson, 2012), Frasnian (Henderson, 1980), Famennian and Early Carboniferous (Henderson, 1987) and middle Carboniferous (330–325 Ma: Black et al., 1979; Bell, 1980; Draper et al., 2004) phases.

The southern mega-shear extends along the Darling River Lineament and subsidiary Cobar–Inglewood Lineament into the offshore Cato Fracture Zone (Figs. 11B, 14 top left). This prominent structure has been identified on structural grounds (Rod, 1966 [Birrie Geosuture]; Leitch and Scheibner, 1987; Scheibner and Veevers, 2000; Li and Powell, 2001; Neef, 2012) and as a gravity (G7) and associated structural lineament (R1-2-L1) (O'Driscoll, 1983; Campbell and O'Driscoll, 1989). Structural offsets indicate dextral displacement (Wopfner, 1960; Rudd, 1961; Rod, 1966; O'Driscoll, 1983; Li and Powell, 2001). Movement along the Darling River Lineament led to dextral displacement of the NNEO with respect to the SNEO and probably of the Drummond Basin with respect to the Lambian Basin (Veevers, 1984, figs. 223, 224; Li and Powell, 2001, figs. 13, 14b). Paleomagnetic data from the Grassmere Knee Zone (Vérard, 2009), at the western Darling River Lineament, indicate dextral movement postdating the Early Devonian (Lochkovian–Pragian) Mount Daubeny Formation. The eastern Darling River/Cobar–Inglewood Lineament intersects the Meandarra Gravity Ridge, with Bouguer gravity data showing no displacement (Krassay et al., 2009). This is to be expected as ASO-related tectonic extrusion ended by middle-late Viséan, well before latest Carboniferous–earliest Permian development of the gravity feature.

The Darling River Lineament can be continued westward into the Menindee and Tar(r)ar(r)a Troughs (Scheibner, 1974; Neef, 2012), and far beyond as part of a global fault zone that may date back to Proterozoic. This fault zone apparently extends westward, within Gondwana, along an Antarctica–Australia intraplate shear zone (Willcox and Stagg, 1990; Glen et al., 1992; Scheibner and Basden, 1998) and/or the Albany Mobile Belt of dextral movement (Harris and Beeson, 1993), into the Narmada–Son Lineament/Central Indian Tectonic Zone (Acharyya, 2003; Naganjaneyulu and Santosh, 2010) of dextral movement (Sen, 1991; Harris and Beeson, 1993), and into eastern Africa (Fig. 15; Reeves et al., 2004).

A subsidiary southern mega-shear, extending along the Lake Blanche Fault Zone and Olepoloko Fault Zone into the Lachlan Transverse Zone, shows a more complex pattern of sinistral and dextral movements (Scheibner and Basden, 1998; Glen and Walshe, 1999; Scheibner and Veevers, 2000; OZ SEEBASE™, 2005; Vérard and

Glen, 2008). The Lachlan Transverse Zone was active from Ordovician to Carboniferous and shows for the Tarana Granite of the Bathurst Batholith a distinctive pattern of eastward-younging middle Carboniferous intrusions (Figs. 11B, 14 top left; Pogson and Watkins, 1998). Interaction during Devonian and Carboniferous between the two southern mega-shears of ENE–WSW and WNW–ESE orientation, forming the Darling Basin Conjugate Fault System, may have contributed to the complex, sequentially compressional and extensional, development of the Darling Basin (Scheibner and Basden, 1998; Neef and Bottrill, 2001; Willcox et al., 2003; Cooney and Mantaring, 2007; Neef, 2012).

The SNEO paleolatitudinal evolution (Figs. 5B, 7D, 14 top right) indicates ending of Australia–Asia convergence during middle-late Viséan, around 330 Ma. Ending coincided with end of sedimentation in the Bundock pull-apart basin and Clarke River Basin (Wyatt and Jell, 1980; Jell and Playford, 1985; Bain and Draper, 1997; Scheibner and Veevers, 2000; Withnall and Henderson, 2012), located in a tensile jog along the northern mega-shear (Fig. 14 top left, C, D), and also with end of sedimentation in the Texas Basin (Jenkins et al., 1993; Scheibner and Veevers, 2000) near the southern mega-shear (Fig. 14 top left, E). End of convergence coincided also with ending of subduction-related volcanism in the eastward-displaced NNEO, whereas volcanism continued in the SNEO (Fig. 14 top left, F, G; Murray et al., 1987; Veevers et al., 1994b; Scheibner and Basden, 1998; Scheibner and Veevers, 2000) till about 305 Ma (Roberts et al., 2006; Cawood et al., 2011b; Jeon et al., 2012). Eastward extrusion and displacement of the NNEO versus the SNEO may have led during Late Devonian–Late Carboniferous to choking of the NNEO subduction zone (Korsch, 2004) and to differentiation in crustal architecture along the length of the NEO (Murray, 1994), with the NNEO thrust over the Thomson Orogen but the SNEO thrust beneath the Lachlan Orogen (Korsch et al., 1997; Korsch, 2004).

The Late Devonian–Early Carboniferous Campwyn Volcanics, Yarrol Terrane, show common presence of inherited zircons with Proterozoic to Middle Devonian ages indicating (reworked?) Proterozoic basement for the NNEO (Bryan et al., 2004). Permo–Carboniferous igneous rocks in contrast show lack of significant zircon inheritance (Allen et al., 1998; Bryan et al., 2004). ASO-related eastward tectonic extrusion through lower crustal ductile flow may explain the presence of inherited zircons in the Late Devonian–Early Carboniferous Campwyn Volcanics, and also in the Drummond Basin succession, the Retreat Batholith of the Anakie Inlier and the Mt Morgan Stratigraphic Assemblage, as well as their absence in Permo–Carboniferous igneous rocks that post-date ASO-related tectonic extrusion. Tectonic extrusion also concurs with Bryan et al.'s (2003) pioneering suggestion that the magmatic and tectonic evolution of the NNEO and its hinterland may have been driven by east-directed far-field forces from the ASO, rather than west-directed Paleopacific subduction. Eastward tectonic extrusion through lower crustal ductile flow also would reconcile with an extensional origin for Late Devonian–Early Carboniferous silicic volcanism in the NNEO, Drummond Basin (Johnson and Henderson, 1991; De Caritat and Braun, 1992; Murray, 1994; Henderson et al., 1998; Champion et al., 2009), Broken River Province and Connors Arch (Bryan et al., 2004), assuming that the basins formed as ramp-basins comparable to ramp-basins/synclines identified in the Adavale Basin (Section 4.4.3).

Evolution of Devonian shorelines and Paleozoic I–S lines provides further support for the ASO Tectonic Extrusion Hypothesis. Devonian shorelines for eastern Australia based on ostracod data indicate large-scale sinistral offset along the CRFZ of the northern mega-shear, although conjugate dextral offset along the southern mega-shears is less clear (Fig. 16A, B; Talent et al., 2000). An I–S line for Early–Middle Paleozoic granitoids indicates sinistral displacement near the CRFZ, but a late Paleozoic I–S line shows no such displacement (Fig. 16C, D; Richards, 1980) as expected for granitoids postdating tectonic extrusion. Two of Powell's paleogeographic maps (Fig. 17; Powell in Veevers, 1984; Li and Powell, 2001) feature eastward displacement of

the NNEO and Thomson Orogen in apparent agreement with the ASO Tectonic Extrusion Hypothesis, although such a cause for displacement was not advanced by Powell. One of the maps, reflecting on the origin of the Texas–Coffs Harbour Oroclines (Fig. 17A; Veevers, 1984), indicates eastward displacement of the NNEO during latest Devonian–earliest Carboniferous. The other map, an Early Carboniferous (Viséan) reconstruction of Australia within Gondwana (Fig. 17B; Li and Powell, 2001), also indicates eastward displacement of the Thomson Orogen and NNEO with sinistral/dextral movements along northern/southern bounding mega-shears.

For veracity of a tectonic extrusion model, approximate mass balance between crustal shortening and crustal extrusion should be demonstrated. For the ASO Tectonic Extrusion Hypothesis there is first-order mass balance between crustal shortening across the Larapintine Graben and crustal extrusion of the Thomson Orogen and NNEO. Powell estimated north–south crustal shortening across central Australia and the TOS at 100 to 150 km. Eastward displacement of the NNEO has been estimated at 200 km or more, both with respect to the Broken River Embayment and Hodgkinson Province in the north (Figs. 11B, 14C; Bell, 1980; Harrington, 1981) and with respect to the SNEO in the south (Rod, 1966; Runnegar, 1974; Leitch, 1975; Murray et al., 1987; Veevers et al., 1994b). Crustal volumes representing 100 km shortening across the width of the “compression box” (Fig. 11A) and 200 km eastward displacement along the length of the NNEO (Fig. 11B), outside the “compression box”, are in approximate mass balance indeed. Further crustal shortening in central Australia, up to 150 km, may account also for southward and northward mass displacements within the internal TOS (Fig. 11B).

4.4.3. Classic Eromanga–Brisbane seismic profile revisited

The Eromanga deep seismic profile (Finlayson, 1990b; Finlayson et al., 2004) through southern Queensland extends from the Adavale Basin eastward, across the Nebine Ridge–Anakie Inlier, into the NEO (Fig. 18A, B). The profile provides detailed reflection and refraction data across the region of purported ASO-related tectonic extrusion. The western section of the profile (Finlayson, 1990a; Finlayson et al., 1990a) is reproduced in Fig. 19A as originally interpreted. This section highlights consistent presence of non-reflective upper crust, highly reflective lower crust, and four major west-dipping, crustal-penetrating, detachment zones. Finlayson et al. (2004) interpreted the minimal topography of the sub-horizontal Moho as relief of stresses in a ductile lower crust at elevated temperatures. The two eastern detachment zones (Fig. 19A; Foyleview, Westgate) sole in the Moho or uppermost lithospheric mantle, and the two western detachment zones (Quilpie Ramp, unnamed) sole near the upper–lower crustal boundary. The detachment zones consistently show eastward ramping-up of highly reflective lower crust, with consequent reduction of non-reflective upper crust. Leven et al. (1990) and Finlayson et al. (1990a) interpreted basin development across the seismic profile as substantial horizontal movement between lower and upper crust. They interpreted the Westgate Trough and Quilpie Trough of the Adavale Basin (Fig. 18B) as a ramp basin and ramp syncline respectively, reflecting more than 35 km westward movement of upper crust over lower crust (Fig. 18C, upper crustal arrow). Such westward relative movement fits the paradigm of eastern Australian tectonic deformation driven by westward subduction of the Paleopacific (Evans et al., 1990; McKillop et al., 2007).

Alternatively, eastward movement of lower crust versus upper crust leads to the same relative movement pattern and fits the hypothesis of ASO-related tectonic extrusion driven by lower crustal ductile flow (Fig. 18C, lower crustal arrow). The Cooper Basin to Nebine Ridge section of the Eromanga reflection profile (Fig. 19A) has been re-interpreted accordingly in terms of eastward lower crustal flow (Fig. 19B), guided by an emerging model for eastward expanding uplift of the Tibetan Plateau (Section 3.5). The re-interpreted profile (B) retains the original seismic compartmentalization (A), but highlights highly reflective lower crust versus non-reflective upper crust. The

re-interpretation also highlights differential movement of lower crust versus upper crust as caused by eastward lower crustal ductile flow, driving tectonic extrusion of lower and upper crust. In addition to the major west-dipping crustal-penetrating detachments—Quilpie Ramp, Westgate Detachment and Foyleview Detachment—the original interpretation also shows an eastern lower crustal segment with substantially higher seismic velocity than more western lower crust. The re-interpreted profile highlights substantial eastward ramping-up of highly reflective lower crust along the Foyleview Detachment over this lower crustal buttress. Eastward ramping-up is shown also for the Westgate Detachment and the Quilpie Ramp, albeit less conspicuously.

The proposed driver for this pattern of lower crustal masses ramping-up eastward over more resistant lower crustal masses is ASO-related tectonic extrusion, facilitated through lower crustal ductile flow and partial melt. Upward guidance of melt along detachment zones and ramps may have led to regionally extensive felsic magmatism in the Warburton Basin and Thomson Orogen (Murray, 1994; Champion et al., 2009), with near-surface formation of granitic plutons of Late Devonian and Early Carboniferous age. This has analogy to upward movement of melts along roof thrusts and formation of High Himalayan leucogranites in southern Tibetan–Himalayan channel flow models (Nelson et al., 1996; Wu et al., 1998; Grasemann et al., 2006; Scaillet and Searle, 2006; Ferré et al., 2012). Roma Granite magma (Finlayson, 1990b, maps 1, 4, Late Devonian–Early Carboniferous [G2 granite]; Murray, 1990, 1994, Early Carboniferous) may have traveled up the Foyleview Detachment. Likewise, Nebine Ridge granitic magma (Fergusson et al., 2009, G2 granite) may have traveled up the Westgate Detachment. Epithermal gold mineralization in the Drummond Basin and Anakie Inlier, dated 350–340 Ma by K–Ar and SHRIMP U–Pb zircon methods (Perkins et al., 1995; Henderson et al., 1998; Champion et al., 2009), may have resulted from accompanying fluid flows. This mineralization has been linked tentatively to a suite of Early Carboniferous paleomagnetic overprints observed across the TOS and parts of the Australian craton (Section 4.5.2; Klotwijk, 1996a, 1998, 2009). Granites above the Quilpie Ramp and the unnamed ramp near the Canaway Ridge (Fig. 19) lack age control. Extensive Middle–Late Carboniferous granites of the Cooper Basin's Big Lake Suite (Gatehouse et al., 1995; McLaren and Dunlap, 2006; Meixner and Holgate, 2009) occur further west, off Fig. 19. Large-scale lenticles, or pod-shapes zones, in the lower crust of the western Adavale Basin (Fig. 18B, Barcoo and Warrabin Troughs; Finlayson et al., 1989) deserve further study regarding their proposed uppermost mantle underplating origin, or alternatively, possible derivation from ductile flow in the lower crust driven by ASO-related tectonic extrusion.

4.4.4. Fold draping onto the Nebine Ridge and Anakie Inlier

Eastward ramping-up of lower and upper crust along the Westgate Detachment onto the Nebine Ridge is re-interpreted for the Eromanga deep seismic profile (Section 4.4.3; Fig. 19B). The Anakie Inlier and flanking western and eastern Drummond Basin represent a northern extension to the Nebine Ridge structure (Fig. 18B; Murray, 1994). Regional mapping of the Drummond Basin (Olgers, 1972) has demonstrated a conspicuous pattern of fold structures, draping onto the northern Nebine Ridge and Anakie Inlier. Fold structures follow gentle to arcuate changes in the overall north–south structural trend, with tight fold draping around the northern edge of the Anakie Inlier through the passage with the Lolworth–Ravenswood Block (Fig. 20C; Olgers, 1972; Johnson and Henderson, 1991; De Caritat and Braun, 1992; Davis and Henderson, 1996; Henderson et al., 1998). Draping of western Drummond Basin fold structures onto and around the Anakie Inlier suggests a pattern of crustal movement that resembles frontal ramping-up from lower crustal levels in the Longmen Shan with oozing of Tibetan crust around the barrier of the Sichuan Basin (Section 3.5; Figs. 9A, 20A, B). Such movement is interpreted as driven by ASO-related tectonic extrusion, transmitted hydraulically through eastward flow of ductile lower crust.

4.4.5. Hydraulic connectivity and tectonic synchronicity

A key aspect of the ASO Tectonic Extrusion Hypothesis is hydraulic transmission of far-field forces through the lower crust of the “compression box”, with re-alignment of force vectors along local pressure gradients. Collision-related Devonian–Carboniferous far-field forces are oriented north–south to northeast–southwest, about perpendicular to the outline of the Larapintine Graben across the “compression box” (Figs. 3, 11). The northern and southern mega-shears that are bounding tectonic extrusion are oriented also at large angles to the far-field forces. Further east, the “free oceanic boundary” of the Paleopacific provides an about east–west pressure gradient where the mega-shears extend toward oceanic crust, with a more pronounced pressure gradient during times of subduction rollback. Analogous to deflection of movement (force) vectors in the Tibetan and Anatolian–Aegean regions (Section 3.2; Fig. 8B), hydraulic connectivity through lower crustal ductile flow deflects the far-field force vectors from southward to eastward. Such deflection of far-field force vectors offers a mechanism for synchronicity of orogenic events with differing tectonic grains (Coward and Ries, 1986, p. xii), in part reflecting simultaneous transmission of differently oriented forces through different crustal levels of differing strengths (Masek et al., 1994; Axen et al., 1998; Vanderhaeghe, 2009). North–south compression from Australia–Asia convergence may have led about simultaneously, to the east–west tectonic grain of the ASO by direct upper crustal transmission of far-field forces, and also to the north–south tectonic grain of the Quilpie Orogeny (Thomson Orogen) and the, in part, east–west tectonic grain of the Kanimblan Orogeny (Lachlan Orogen) by hydraulic transmission through ductile lower crust of far-field forces, re-aligning along local pressure-gradients (Fig. 11B).

A different force pattern is expected outside the “compression box” in regions where Australia–Asia suturing had not progressed yet, such as the northwestern Australian craton during Early Carboniferous (Fig. 11A). The Canning and Bonaparte Basins here would not directly experience compressional forces from Australia–Asia convergence, during all but the final stage of convergence when Australia–Asia suturing had expanded over a front wider than the central New Guinean cratonic promontory. Instead, the two basins may have experienced extensional forces, derived from local misalignment between far-field force vectors and fault boundaries of the “compression box” (Braun et al., 1991), from leakage of compressional forces across the fault boundaries of the “compression box”, and from far-field extensional conditions in regions where continental convergence had not yet progressed to continental suturing. With ongoing convergence and collision, progressive suturing expanded the width of the “compression box”, changing local extensional conditions into wider compression.

Compilation of Devonian and Carboniferous tectonic phases (5) for selected regions within (6) and without (2) the “compression box” (Fig. 13) indicates Australia-wide connectivity between compressional and extensional deformation phases. Within rather broadly defined time intervals during Late Devonian and Early Carboniferous, there is remarkable coincidence between three phases of compressional deformation within the “compression box” and three phases of extensional deformation without the “compression box”, collectively known in the Canning Basin as the Pillara Extension (Shaw et al., 1994; Braun and Shaw, 1998). That is the Henbury Movement and Tabberabberan Orogeny (383–377 Ma) versus the Gogo Extensional Movement, the Pre-Brewer/Kerridy Movement (365–362 Ma) versus the Van Emmerick Extensional Movement, and the Brewer and Kennedy Movements and Kanimblan Orogeny (355–340 Ma) versus the Red Bluffs Extensional Movement. A mechanism for such synchronicity between compressional deformations and extensional movements is bound to be conjectural (Braun et al., 1991; Braun and Shaw, 1998). Hydraulic connectivity across the western border of the “compression box” is an option to be explored further, in conjunction with far-field extensional stress from the western, not-yet-sutured, region of continental convergence. Subsequent to this triple compression–extension dichotomy, compressional

deformation expanded Australia-wide during middle(–Late?) Carboniferous (333[–315?]Ma: Meda Compressional Movement in the Canning Basin; Alice Springs and Mount Eclipse Movements; mega-kinking across the TOS; mid-Carboniferous north–south deformation in north Queensland (Bain and Draper, 1997; Champion et al., 2009)), indicating that Australia–Asia suturing had progressed beyond the cratonic promontory of central New Guinea, effectively expanding the “compression box”. An illustrative analog for this Paleozoic compression–extension dichotomy is the current Arabia–Eurasia convergence, with simultaneous compression along the Bitlis–Zagros zone of collision and extension along the Hellenic and Cyprian Arcs (Allen, 2010) to the east and west respectively of the Dead Sea Fault Zone.

4.5. Implications

4.5.1. Post-tectonic extrusion – Kiaman – deep crustal remagnetization

Collision of Australia/northeastern Gondwana with central Asia ended by middle-late Viséan, around 330 Ma (Figs. 5B, 6, 21B), just before the Permo-Carboniferous Reverse Superchron (PCRS or Kiaman). This period of continuous reverse polarity of the earth's magnetic field lasted about sixty million years (Fig. 21A). Start of the PCRS was dated by Opdyke et al. (2000) between 316 and 318 Ma, based on integrated magnetostratigraphic and SHRIMP U–Pb zircon study of ignimbritic successions in the western Tamworth Belt, SNEO. Roberts et al. (2003) revised the 316–318 Ma start date to 327 Ma in a heuristic correction for homogeneity problems with the SL13 calibration standard, but this awaits corroboration (see Gradstein et al., 2012). The poorly defined end of the PCRS is commonly estimated at 265 Ma (Menning, 1995; Idnurm et al., 1996; Gradstein et al., 2012). Heated, ASO-related tectonically extruded lower and upper crust thus began cooling at the start of an exceptionally long period of continuous reverse polarity of the earth's magnetic field, presenting optimal conditions for uniquely reverse overprinting when magnetization carriers cooled through their blocking temperatures.

Following discovery of highly magnetized lithosphere on Mars in the absence of a Martian magnetic field, it was realized that hematite can contribute significantly to remanent magnetization of the earth's deeper crust (Kletetschka et al., 2000a; Dunlop and Kletetschka, 2001). Further studies have shown that multidomain hematite (Kletetschka et al., 2000b), slowly cooled fine lamellae of hematite–ilmenite intergrowth (Robinson et al., 2002; Kasama et al., 2004; Harrison et al., 2006; McEnroe et al., 2007; Fabian et al., 2008), and hematite with fine intergrowths of magnetite and/or maghemite (Schmidt et al., 2007) can retain remanent magnetizations with intensities comparable to magnetite. These hematite magnetizations can be very stable. With unblocking temperatures up to a hundred degrees higher than magnetite such hematite carriers can contribute a significant magnetic signal from crustal levels substantially deeper than for lowermost magnetite contributions (Kletetschka et al., 2002; McEnroe et al., 2004, 2009). How deep depends on the paleo-geothermal gradient, but hematite likely can make a substantial remanent contribution at lower crustal or uppermost lithospheric mantle levels (Ravat et al., 1992; Purucker, 2007; Dunlop et al., 2010; Purucker and Clark, 2011). Lower crust, tectonically extruded before, and cooled during, the PCRS may thus contribute a reverse polarity component to the lithospheric magnetic field. Conversely, the extent of ASO-related lower crustal outflow may be traceable by mapping such reverse polarity re-magnetization (Betts et al., 2007).

Australian lithospheric magnetic anomaly maps, based on satellite (POGO, MAGSAT, ØRSTED, CHAMP) or aeromagnetic data, are prospective for delineating deeper crustal magnetization components. Anomaly maps derived over the past three decades from satellite scalar and scalarized vector data have consistently shown prominent negative magnetic anomalies. The outlines of these negative anomalies are oriented westnorthwest–east-southeast along the cratonic Larapintine Graben and north–south along the Thomson and internal Lachlan Orogens of the TOS (Fig. 21C–H; Purucker et al., 1997; Ivers et al., 2000; Sabaka

et al., 2004; Maus et al., 2008; Lesur et al., 2008). This pattern of negative magnetic anomalies is also well visible in the long wavelength component of the aeromagnetic map of Australia (Fig. 21i; Milligan and Franklin, 2004; Milligan et al., 2004, 2010). The negative signature of these prominent magnetic anomaly structures might relate merely to (i) low content of magnetic carriers and/or (ii) crustal temperatures above their unblocking temperatures, assuming data presentation bias by leveling errors. However, obvious correspondence between the anomaly structure of all maps derived from four different satellite missions over the past three decades and also the long wavelength component of the aeromagnetic map makes it unlikely that the negative signature of the anomaly structures merely represents leveling bias. The negative magnetic anomalies therefore may represent (iii) induced magnetization and/or (iv) reverse polarity remanent magnetization structures. Induced magnetization (iii) is not an option for the north–south anomaly in the internal TOS, but could have been for the east–west anomaly along the Larapintine Graben (D. Ravat personal communication 2011) if not negated by persistence of negative anomaly features after reduction to the pole (Fig. 21i). The preferred interpretation option therefore is a reverse polarity remanent magnetization component of deeper crustal origin and of Kiaman (PCRS) age (iv). Conclusive differentiation between the four source options (i–iv) may have to await future satellite vector data rather than current scalar and/or scalarized vector data.

The concept of post-Variscan, Kiaman, remagnetization of deeper crust is not new and was proposed for a magnetic low observed from MAGSAT data over the type region of the Variscan Orogen in central Europe (Pucher and Wonik, 1997; Taylor and Ravat, 1997). Beside deeper crustal remagnetization, widespread mid Carboniferous to Late Permian paleomagnetic overprinting was also described for the Bohemian Massif (Reisinger et al., 1994) and other Variscan massifs—the Vosges, Black Forest, Massif Central (Edel, 2001) and the Armorican Massif (Edel and Aifa, 2001). This complements the deeper crustal remagnetization analogy between European Variscan massifs and ASO-related tectonic extrusion with an uppermost crustal paleomagnetic overprinting analogy, explored further hereinafter (Section 4.5.2).

The negative magnetic anomalies along the Larapintine Graben and the internal TOS were identified by Wellman (1995, 1998), from an early version of the Australian aeromagnetic map, as a domain of lesser magnetic crust (Fig. 22A). Wellman also identified bands of magnetically reworked crust, bordering the lesser magnetic domain and crosscutting an eastern higher magnetic domain (Fig. 22B). If the lesser magnetic domain of the internal TOS represents a Kiaman reverse polarity overprint, then the higher magnetic domain of the external TOS may reflect a normal polarity overprint from a Late Cretaceous thermal event (105–95 Ma; Lackie and Schmidt, 1996; O'Sullivan et al., 1999a; Klootwijk, 2009) that preceded opening of the Tasman Sea (85 Ma; Gaina et al., 1998; Müller et al., 2000; Norvick et al., 2001). This thermal event occurred about midway during the Cretaceous Normal Polarity Superchron (CNPS: Idnurm et al., 1996, 115–83 Ma; Gradstein et al., 2012, 126.3–83.6 Ma). Subsequent cooling may have imparted a normal polarity overprint affecting the southeastern seaboard of Australia and the EARS region previously tectonized during latest Carboniferous–Early Permian (Fig. 22B). Joint presence in far eastern Australian crust of a Late Paleozoic, Kiaman, reverse polarity overprint and a Late Mesozoic, CNPS, normal polarity overprint would complicate magnetic identification of the easternmost extent of Middle–Late Paleozoic lower crustal flow.

4.5.2. ASO-related paleomagnetic overprinting

Pervasive Middle–Late Paleozoic paleomagnetic overprinting in Australia has been recognized for long, in particular from early studies in the Amadeus Basin, Flinders Ranges, Lachlan Orogen and TOS (McWilliams, 1977; Kirschvink, 1978; Klootwijk, 1980; Goleby, 1981; Schmidt and Embleton, 1987). Overprints in the Amadeus

Basin and Flinders Ranges showing broad association with Middle–Late Paleozoic segments of then prevailing pole paths, were interpreted collectively as “Alice Springs Orogeny” (ASO) overprints. Such overprints are directionally quite dispersed, being of pre-tectonic, syn-tectonic, or post-tectonic origin, and of mixed polarity. With further definition of Middle–Late Paleozoic pole paths, the KG and SNEO paths in particular (Klootwijk, 2010), it has been realized that overprints close to the latest Carboniferous–Early Permian segment of these pole paths (Fig. 23B) may have been misinterpreted as ASO-related. Such overprints are more likely associated with a subsequent phase of Australian tectonism, starting in Stephanian (~305 Ma; Figs. 5A, 23A; L3 loop; Klootwijk, 2009, fig. 4) and driven by a major movement change of Gondwana versus Laurasia within Pangea (Section 4.5.6). This change in global plate movements led to formation of the EARS, the Bowen–Gunnedah–Sydney basin system, oroclinal deformation of the SNEO, and also to further paleomagnetic overprinting in tectonically weakened zones such as the Larapintine Graben. It has been realized also that a substantial set of overprints from mainly the Flinders Ranges, and less so the Amadeus Basin, originally identified as Delamarian overprints (Klootwijk, 1980), could be re-interpreted, with swapped polarity, as middle Carboniferous ASO overprints (Fig. 23A, B; L2 loop; Klootwijk, 1996a). Original interpretations of ASO and Delamarian overprints should be treated with caution.

Many paleomagnetic studies, in the Larapintine Graben and TOS in particular, have described primary and/or overprint magnetizations that can be re-interpreted—guided by comparison with the KG pole path (Klootwijk, 2009, fig. 2), its upgrade (Fig. 23B), and the Carboniferous SNEO pole path (Figs. 5A, 23A)—as overprints of Devonian and in particular Early-middle Carboniferous origin (Supplementary Table 3). These overprints, acquired at pre-, syn- and post-tectonic stages of the ASO, may have a unifying cause in Australia–Asia collision-related tectonism, specifically in upper crustal hydrothermal fluids derived from ASO-related lower crustal ductile flow. Mapping of these overprints may outline the regional extent of ASO-related lower crustal flow and tectonic extrusion and may locate fluid pathways facilitating mineralizations.

The dilemma of Australian Middle–Late Paleozoic pole path interpretations (Klootwijk, 2009, 2010) may come closer to a solution with recognition that ASO-related tectonic extrusion led to a dichotomy in ages of magnetic overprinting at uppermost and at deeper crustal levels. Paleomagnetic overprints in the uppermost crust became acquired largely synchronous with Variscan tectonism, with hydrothermal fluids cooling to low ambient temperatures close to the earth's surface. In contrast, thermal inertia at deeper crustal levels delayed remagnetization till post-tectonic, Kiaman, times.

4.5.3. Centralian gravity anomalies and crustal stress

Central Australia is characterized by profound gravity anomalies (Fig. 24A; Lambeck, 1984; Braun et al., 1991; Morse et al., 1992; Bacchin et al., 2008; Aitken et al., 2009a), aligned east–west across the Georgina, Ngalia, Amadeus and Officer Basins and the Arunta and Musgrave basement blocks, ranging from +40 mGal to –150 mGal (Bouguer). The anomalies reflect a protracted tectonic evolution of central Australia, from Late Paleoproterozoic (Myers et al., 1996; OZ SEEBASE™, 2005; Aitken et al., 2009a) till the Late Paleozoic ASO. Survival of these substantial gravity anomalies over 300+ myr following the ASO, indicates a longstanding and ongoing capability of crust and lithospheric mantle to simultaneously maintain a substantial isostatic dis-equilibrium and a mechanical equilibrium (Beekman et al., 1997).

Could such a mechanical equilibrium be maintained over more than 300 myr by the balance of forces acting on the margins of the Australian plate (Sandiford et al., 1995; Coblentz et al., 1998)? This is unlikely; it would require a continuing north–south stress pattern across central Australia. The present Australian stress pattern is complex (Zhao and Müller, 2003; Burbidge, 2004; Sandiford et al., 2004; Hillis et al., 2008; Braun et al., 2009). Highly localized observations are dominated by north–south stresses in the Amadeus Basin and

east–west stresses in the Cooper–Eromanga Basin (Reynolds et al., 2005). In addition there is a sprinkling of directionally variable stresses across central Australia. The current east–west stress pattern in the Cooper–Eromanga Basin is difficult to explain by plate boundary forces. Its origin is enigmatic (Hillis et al., 1998), but could reflect tectonic extrusion-related crustal structural fabric (Section 4.4.1). Noteworthy is spatial correspondence between a northeastern Australian region of near-isotropic stress (Reynolds et al., 2002; Fig. 24B) and the region primarily sculpted by ASO-related tectonic extrusion (Fig. 11B), also characterized by low strain rate and low seismicity (Burbidge, 2004). Would this very low stress and strain regime merely reflect selective incorporation of present-day boundary forces within the modeling, or might it reflect ASO-related tectonic extrusion? More specifically, could Middle–Late Paleozoic lower crustal ductile flow toward minimal pressure regions have created a long-time persistent quasi-isotropic stress regime in basement at the outer region of tectonic extrusion?

The present, predominantly north–south, stress pattern of the Amadeus Basin reflects Australia's Late Neogene collision with the Sunda Platform and its Neogene collision with island arcs across the cordillera of central New Guinea (Klootwijk, 2009; Hall, 2011). It is unlikely that substantial north–south stresses would have persisted during Mesozoic and Early Cenozoic as there is no indication for north–south collisional tectonics across the northern plate boundary during that period, although ridge push from the Neotethyan Whar-ton Ridge may have led to north–south stress patterns as tentatively modeled for mid Cretaceous. If representative, such patterns changed fundamentally with Late Cretaceous (95–90 Ma) opening of the Tasman Sea leading to more rigorously defined Eocene and Miocene stress patterns that are widely different from the present pattern (Dyksterhuis et al., 2005; Dyksterhuis and Müller, 2008; Müller et al., 2012).

Stress modeling of the Australian plate based on boundary forces and topography has generally shown low intensity, spatially variable, stresses (Stephenson and Lambeck, 1985; Coblentz et al., 1998; Zhao and Müller, 2003), unlikely to have contributed significantly to maintain mechanical equilibrium of a crust with prominent gravity anomalies (Beekman et al., 1997; Aitken et al., 2009b). This indicates that the central Australian crust maintained significant strength over the past 300+ myr. Prior existence of such crustal strength, over the duration of the ASO in particular, would counter arguments for tectonic extrusion through ductile flow in the lower crust. However, Sandiford and co-workers' studies indicate that such obvious crustal strength would have been an *expected outcome* of ASO tectogenesis, rather than an *impediment* (Sandiford and Hand, 1998; Sandiford et al., 1998; Sandiford, 1999; Sandiford et al., 2001; Sandiford and McLaren, 2002, 2006). Their arguments for progressive strengthening of the crust focus on thermal evolution of crust containing abundant HHGGs, blanketed by sediments and progressing through graben-forming and inversion processes. The present hypothesis of tectonic extrusion through lower crustal ductile flow, and perhaps partial melt, may add a further mechanism for crustal strengthening—more mobile, heat producing and crustal weakening, elements would have been extruded first (Sandiford and McLaren, 2002; Aitken et al., 2009a; Vanderhaeghe, 2009; Lexa et al., 2011), leaving a drier crust with reduced mobility (Yamasaki and Stephenson, 2009). Strengthening of lower crust that is cooling following ductile flow, has been reported also for well-studied, exhumed, continental lower crust elsewhere (Klepeis et al., 2007; Aitken et al., 2009b; Dumond et al., 2010).

4.5.4. Namurian lacunae and Kanimblan Highlands

Denudation over the 100+ myr lifespan of the extended ASO has resulted in some 20 km of exhumation (Veevers, 2009). Piedmont-type deposition in central Australia has been associated with major phases of denudation around 450, 400, 360 and 330 Ma, and an ASO source has been suggested for Middle Devonian–Early Carboniferous Drummond Basin and Lambian Basin sedimentation (Fig. 25; Li and Powell, 2001;

Glen, 2005; Champion et al., 2009). Late Carboniferous channeling of sediments to the New England and Canning/Bonaparte cratonic margins has been documented (Fig. 17; Powell in Veevers, 1984; Fig. 25; Li and Powell, 2001). However, dearth of deposits closer to the ASO source in central Australia, for middle Carboniferous (~330 Ma, paroxysmal phase) and Namurian (Namurian lacunae) in particular, has been baffling. It led Powell and Veevers (Veevers, 1984; Powell and Veevers, 1987; Veevers et al., 1994c; Li and Powell, 2001) to postulate a Namurian continental glacial ice sheet across eastern and central Australia, limiting sedimentation to the NEO in the east and the Canning and Bonaparte Basins in the west. It has been demonstrated subsequently that Late Carboniferous glaciation in Australia was limited to alpine glaciers and that continental ice sheets formed during Permian only (Isbell et al., 2003; Jones and Fielding, 2004; Fielding et al., 2008; Isbell et al., 2012). Veevers (2006) conceded that Powell and Veevers' original argument for a Namurian continental ice sheet acting as a conveyor agent, was not maintainable as a realistic cause for the Namurian lacunae. So what could have caused this remarkable hiatus across Australia that has a stratigraphically more extensive counterpart across Gondwanan India (Wadia, 1953; Gansser, 1964; Krishnan, 1968; Ibrahim Shah, 1977)?

The ASO Tectonic Extrusion Hypothesis offers a plausible alternative for the Namurian lacunae. Upon Australia–Asia collision, cratonic Australia was deformed within the confines of the “compression box” with uplift through thermal expansion associated with ductile flow of heated lower crustal, from the Larapintine Graben and the TOS, toward and as far east as the EARS boundary of the “compression box”. Uplift likely progressed outward from the central Larapintine Graben. The “compression box” highlands became a region of active erosion but limited sedimentation, with the eroded sediment load transported to lowlands outside the “compression box”. Powell (Veevers, 1984) (Fig. 17A), Scheibner and Veevers (2000, figs. 251, 257, 258), and Li and Powell (2001) (Fig. 25) show sediment deposition by major transporting rivers near the eastern, coastal, sections of the northern and southernmost mega-shears of the tectonic extrusion model (Fig. 11B). From provenance study of NEO sediments, Korsch et al. (2009a) concluded likewise a major eastward sediment outlet through the Broken River region, in the vicinity of the northern mega-shear, and another outlet about a hundred kilometers further south, depositing sediment from a northward flowing river through the northwestern Drummond Basin (Olgers, 1972; Murray, 1986).

In addition to upper crustal denudation there was lower crustal tectonic erosion. Heated lower crust was moved by ductile flow, probably with partial melt and fluid flow, from the Larapintine Graben, eastward and southward through the TOS, toward the margin of the “compression box”, where lower crustal thickening and thermal expansion led in Late Carboniferous to the Kanimblan Highlands. The highlands probably extended up to the eastern margin of the “compression box”, as far east as the developing EARS (Scheibner and Basden, 1998, fig. 18; Scheibner and Veevers, 2000, fig. 263; Veevers, 2000, fig. 318A; Geoscience Australia Palaeogeographical Atlas). Post-ASO thermal relaxation progressed at variance. The HHGG Big Lake Suite granites of the Cooper Basin and the Bathurst Batholith of the Eastern Lachlan Orogen intruded during mid-Carboniferous, were uplifted, eroded, subsided and became regions of deposition between early Late and latest Carboniferous (Gravestock and Jensen-Schmidt, 1998; Pogson and Watkins, 1998; Scheibner and Basden, 1998; Veevers, 2000; Draper and McKellar, 2002). Uplift led to surface weathering and alteration of the Early Carboniferous Roma Granite prior to Late Carboniferous–earliest Permian extrusion of the Combargno Volcanics (Murray, 1994), and to middle to Late Carboniferous removal of more than 3 km from the Adavale Basin sedimentary pile (McKillop et al., 2007). Subsidence started shortly after cessation of ASO-related tectonic extrusion. Thermal relaxation is regarded as the major driver for regional sag of Australian cratonic coal basins (Galilee, Cooper, Ackaringa, Pedirka, Officer and Canning) during mainly the Permian, with geo-history plots suggesting cooling following Carboniferous heating and tectonism (Middleton and Hunt, 1989). In

contrast, the easternmost Kanimblan Highlands remained emerged for the major part of the Permian (Veevers et al., 1994c; Scheibner and Basden, 1998; Veevers, 2000; Geoscience Australia Palaeogeographical Atlas), perhaps for thermal interaction with the developing EARS and Bowen–Gunnedah–Sydney basin system.

Increased heat flux from tectonic extrusion-related lower crustal flow may shed new light on a longstanding controversy between fission-track specialists and geomorphologists. Fission track thermo-tectonic imaging of the Australian craton and TOS shows no ages older than the ASO, with widespread annealing of fission tracks testifying for widespread thermal–chemical effects of the ASO (O'Sullivan et al., 1999b; Spikings et al., 2001; Gleadow et al., 2002; Kohn et al., 2002; Tingate and Duddy, 2002). Heat flux from inflowing, ASO tectonic extrusion-related, heated lower crust would have increased the local paleo-geothermal gradient, complicating interpretation of fission track data. Fission track studies generally apply a constant paleo-geotherm. This may have led to overestimation of depth of burial and to conflict with paleo-geomorphologists who may not see evidence in the landscape evolution for purported large-scale denudation (Roach, 2000; Kohn et al., 2003; Leaman, 2003).

4.5.5. Geothermal energy potential

Exploration for geothermal energy depends heavily on high crustal temperatures. Geothermal drilling currently has a practical depth limit around 5 km. Mapped temperature estimates at that depth (Fig. 24C; Chopra and Holgate, 2005; Geoscience Australia, 2007; Budd et al., 2008; Gerner, 2010) rely mainly on borehole temperature data, from oil and gas wells and from water wells in the Artesian Basin (Habermehl and Pestov, 2002). These wells are concentrated in Australian onshore and offshore sedimentary basins (Fig. 24D; OZ SEEBASE™, 2005) and the indicative value of crustal temperature maps is restricted to such well-sampled regions. The maps identify the Cooper–Eromanga Basin as a prime geothermal exploration target, with estimated temperatures at 5 km depth widely in excess of 250 °C. The Cooper–Eromanga upper crustal heat anomaly coincides with large negative magnetic anomalies (Figs. 21, 24F) and with the tectonically extruded, eastward displaced, crustal segment made up of Thomson Orogen and NNEO (Section 4.3.2; Fig. 11B). The nature of this relationship remains to be established.

The Cooper–Eromanga heat anomaly represents an enormous potential resource (Budd et al., 2008; Radke, 2009; Champion et al., 2012). Its vast significance will become clearer with improved understanding of the origin of its heat source and potential for heat replenishment. There are many questions to be addressed through further exploration. Could the heat anomaly reflect enhanced transfer of mantle heat in regions of the Australian continent with reduced crustal thickness (Fig. 24D, E; OZ SEEBASE™, 2005)? Alternatively, if of crustal origin, could this vast heat anomaly reflect wide presence of ASO-related granitoids beneath the Cooper Basin, such as the Big Lake Suite granites (325 Ma; Gatehouse et al., 1995; Gravestock and Jensen-Schmidt, 1998; Meixner and Holgate, 2009) currently explored for geothermal purposes? Would such wide presence be comparable in time and origin to late Viséan–early Namurian (340–325 Ma) swamping of the European Variscan Orogen with S-type granites that probably originated through radiogenic heating (Zwart and Dornsiepen, 1980; Henk et al., 2000)? If the Cooper–Eromanga heat anomaly reflects HHGGs indeed, what then is their wider presence in the Warburton Basin (Gatehouse et al., 1995) and could these HHGGs have been derived from Paleo-Mesoproterozoic HHGG protoliths that abound in the Central Australian Heat Flow Province (Fig. 10B, C; Cull and Conley, 1983; Morgan, 2000; Neumann et al., 2000; McLaren et al., 2003; Minty et al., 2010)? Is there wider presence throughout the Thomson Orogen of anatectic granites derived from lower crustal ductile flow that are enriched in heat-producing radioactive elements (Sandiford and McLaren, 2002; McLaren et al., 2003, 2005; Meixner and Holgate, 2009)? Would partial melting

and crystal fractionation of Paleo-Mesoproterozoic HHGG protoliths and ASO-related tectonic extrusion have led to the remarkable enrichment of radioactive elements, U, Th and K, in Carboniferous granites that are locally abundant throughout the eastern TOS between the southeastern Lachlan Orogen and the Georgetown Block (Champion et al., 2012)? Could this spread of enriched Carboniferous granites be indicative for the easternmost reach of ASO-related lower crustal ductile flow?

Extracted heat can be replenished long-term by conductive heat transfer, but is there potential for short-term heat replenishment by advective/convective fluid transfer from deeper crustal levels (Cox, 2002), along crustal-penetrating detachments such as detailed on the Eromanga seismic profile (Fig. 19; Finlayson, 1990a; Finlayson et al., 1990a)? If so, could crustal permeability have been enhanced during Neogene tectonic re-activation (Sections 3.2, 4.5.3; Gallagher, 1987; Middleton and Hunt, 1989; Gravestock and Jensen-Schmidt, 1998; McLaren et al., 2006) of Middle–Late Paleozoic, ASO-related, crustal structures and stress patterns (Fig. 12G–I)? What is the significance for advective/convective heat transfer of findings of: highly over-pressured brines in Cooper Basin granites; probable mantle-degassing origin of the high CO₂ content in Cooper Basin gas (Wycherley et al., 1999); mantle signature of helium isotopes from Great Artesian Basin sites (Torgersen et al., 1992; Bethke et al., 1999); and large-scale electrical conductivity anomaly structures detailed from magneto-telluric studies (Fig. 24E; Chamalaun and Barton, 1993; Chamalaun et al., 1999; Lilley et al., 2003; Selway et al., 2006)? Could these findings be indicative for a fluid-permeable crust, as interpreted for the Basin and Range Province of western North America (Hilton, 2007; Kennedy and van Soest, 2007), with potential for heat replenishment from deeper lithospheric levels?

4.5.6. Development of SNEO oroclines

Mode and time of orocline development in the SNEO remain controversial. The northern Texas and Coffs Harbour megafolds are widely accepted as a double orocline (Korsch, 1981; Flood and Fergusson, 1982; Fergusson and Flood, 1984; Li et al., 2012b), but veracity of the southern Manning Orocline (Cawood, 1982; Cawood and Leitch, 1985; Korsch and Harrington, 1987; Leitch, 1988; Glen, 2005) has been questioned regarding importance of rotation versus northward translation of the Hastings Blocks (Collins et al., 1993; Roberts and Geeve, 1999; Geeve et al., 2002; Johnston et al., 2002; Offler and Foster, 2008). The presence of a fourth orocline, the eastern Nambucca Orocline, has been suggested (Rosenbaum, 2010, 2012). Its reality and evolution have been tentatively documented by substantive SHRIMP U–Pb zircon data from Early Permian granitoids of the Bundarra and Hillgrove Suites (Rosenbaum et al., 2012), warranting further structural and paleomagnetic testing. A combined triple orocline origin for the northern Texas–Coffs Harbour Oroclines of dextral vergence and the southern Manning Orocline of sinistral vergence has been proposed (Cawood and Leitch, 2006; Klotwijk, 2009; Pisarevsky et al., 2010; Cawood et al., 2011a, 2011b) and questioned regarding coupling of vergences (Glen, 2005; Offler and Foster, 2008). However, coupling of oroclinal structures with different vergences is well established for guirlande structures of the western Alps and Carpathians, formed during Cenozoic convergence of Africa and Europe across the western Neotethys (Fig. 8A; Tapponnier, 1977; Van der Linden, 1985), and for the European Variscan Orogen, formed during Late Paleozoic Gondwana–Laurasia convergence across the Rheic Ocean (Matte, 2001; Weil and Gutiérrez-Alonso, 2007; Corsini and Rolland, 2009; Weil et al., 2010).

Opinions differ widely on mode of formation of the SNEO oroclines. Some models focus on upper crustal involvement (Runnegar, 1974; Korsch and Harrington, 1987; Korsch et al., 1990; Glen, 2005), but most models concern whole crust/lithosphere involvement, coupled with some form of movement of the SNEO or NNEO individually or combined as the NEO (Korsch and Harrington, 1987; Korsch et al., 1990). Focus of these models is on southward movement of the NNEO with respect to the Thomson Orogen, along hypothetical

northwestern (Longreach–Roma[–Lismore]; Evans and Roberts, 1980) to north-northwestern (Gogango–Baryugil; Veevers, 1984; Murray et al., 1987; Fergusson and Leitch, 1993; Veevers et al., 1994b; Veevers, 2000) fault zones ending at the northern oroclines of the SNEO, or along offshore fault zones (Offler and Foster, 2008; Li et al., 2012b). Such models leave the SNEO pinned to the Lachlan Orogen. In contrast, a recent model by Cawood and co-workers (Cawood and Leitch, 2006; Pisarevsky et al., 2010; Cawood et al., 2011a, 2011b) invokes northward movement of the SNEO with respect to the Lachlan Orogen across an uppermost lithospheric mantle detachment, while leaving the NNEO pinned to the Thomson Orogen. The model introduced hereinafter (Fig. 26) likewise involves a pinned NNEO and northward movement of the SNEO. Pinning of the NNEO is supported by geochemical evidence for contiguity of the latest Carboniferous Urannah Igneous Complex and Bulgonunna Volcanics across the Millaroo Fault Zone (Allen et al., 1998) and by paleomagnetic analyses (Fig. 23; Klotwijk, 2009; Pisarevsky et al., 2010; Cawood et al., 2011b).

Opinions on time of formation of the SNEO oroclines vary from Late Carboniferous to Late Permian–Early Triassic. The two recent models favoring northward displacement of the SNEO agree on overall development during Early Permian (latest Carboniferous to early Late Permian), but differ in timing the main phase of oroclinal deformation. Cawood et al. (2011a, 2011b) argue on mainly geological grounds for a *late main phase* of oroclinal bending, at the start of the Hunter–Bowen Phase of the Gondwanide Orogeny (Aldebaran Phase, 270–265 Ma, early Late Permian), with possibly minor rotation at the start of the Tableland Phase of the Gondwanide Orogen (305–295 Ma, latest Carboniferous–earliest Permian maybe extending into Early Permian). In contrast, Klotwijk and coworkers argue on mainly paleomagnetic grounds for an *early main phase* of oroclinal bending. Data from the Myall Blocks (Klotwijk, 2009) indicate a moderate counterclockwise rotation prior to extrusion of the Early Permian Alum Mountain Volcanics (274.1 ± 3.4 Ma; Roberts et al., 1995a), thus preceding the Hunter–Bowen Phase. Further constraints come from reverse polarity, Kiaman, overprints from the Texas, Coffs Harbour and Rouchel Blocks. These are of post-tectonic origin, group on the Early Permian segment of the KG pole path (Aubourg et al., 1994; Klotwijk, 1996b, 2009) and indicate prior oroclinal bending. Primary paleomagnetic data from Alum Rock Conglomerate volcanics of the Texas Block (Aubourg et al., 1994, 2004; Klotwijk, 2009) indicate considerable progress of oroclinal bending at the time of extrusion (292 Ma; Roberts et al., 1996).

The ASO Tectonic Extrusion Hypothesis offers a novel view on evolution of the SNEO oroclines. It shares with the model of Cawood et al. (2011a, 2011b) northward displacement of the SNEO versus the Lachlan Orogen with the NNEO pinned to the Thomson Orogen, and also shares with the model of Li et al. (2012b) an early phase of eastward displacement of NNEO versus SNEO. It differs however, in linking development of the oroclines with latest Carboniferous to Early Permian development of the EARS and Bowen–Gunnedah–Sydney basin system (Fielding et al., 2001; Korsch, 2004; Korsch et al., 2009b), rather than with development of a shear fault to the west and northeast around the SNEO (Cawood et al., 2011a, 2011b). It also differs in driving mechanism, relying on changes in rotation of Gondwana (Figs. 23, 26) rather than directional changes in Paleopacific subduction (Cawood et al., 2011b; Li et al., 2012b).

The novel, tectonic extrusion-related, view is illustrated through several Gondwana relocations (Fig. 26) based on selected paleomagnetic data for the SNEO (Fig. 5A; Supplementary Table 1), highlighting well-dated rotational and latitudinal movements of Australia/Gondwana during Carboniferous and Early Permian. The Early Carboniferous relocation (350 Ma; Fig. 26A) shows Australia moving north, crossing the equator to converge with the CAOB. Tectonic extrusion is in progress, culminating by middle-late Viséan in about 200 km eastward displacement of the NNEO with respect to the SNEO. From then onward, during

Late Carboniferous (Fig. 26B), Australia moves very fast toward south polar latitudes. During latest Carboniferous up to about 306 Ma, Australia rotates counterclockwise (Fig. 26C). A large mega-shear along the Gondwanan margin of the Paleopacific (Visser and Praekelt, 1998; Klootwijk, 2009, fig. 13) develops in eastern Australia as the nascent EARS (Korsch, 2004; Korsch et al., 2009b). This mega-shear leads to development of the Bowen–Gunnedah–Sydney basin system and separates the SNEO ribbon continent from the Lachlan Orogen. By middle-late Stephanian (~305 Ma; Fig. 26C, D; Klootwijk, 2009; see also Cawood et al., 2011a, 2011b) global plate movements change dramatically with Australia/Gondwana starting large-scale clockwise rotation and with initial development of widespread Gondwanan glaciations (Gzhelian–early Sakmarian, ~305–290 Ma; Isbell et al., 2012). The clockwise rotation compresses the eastward-extruded buttress of the NNEO against the SNEO and telescopes the SNEO ribbon. This leads to formation of the Texas–Coffs Harbour and Manning Oroclines as sketched in the Early Permian reconstruction (290 Ma; Fig. 26D), in accordance with paleomagnetic dating control allowing for contemporaneous development of the northern and southern SNEO structures as a triple orocline (Klootwijk, 2009).

Formation of the SNEO oroclines as large-scale buckling of a ribbon continent (Johnston, 2008; Johnston and Gutiérrez-Alonso, 2010) is comparable in timing and geodynamic driver to latest Carboniferous–earliest Permian formation of the Cantabrian–Asturian Arc (CAA) and the Central Iberian Arc (CIA) upon Stephanian establishment of dextral shear between northwestern Gondwana and southwestern Laurasia (von Raumer et al., 2009; Nance et al., 2010; Pereira et al., 2010; Martínez Catalán, 2011b; Shaw et al., 2012). Analyses of paleomagnetic data (Weil et al., 2010, 2012), U–Pb zircon ages and geological data (Gutiérrez-Alonso et al., 2011) and joint patterns (Pastor-Galán et al., 2011) constrain two-stage oroclinal bending of the CAA to middle Moscovian–Gzhelian (310–295 Ma). Formation of the CAA was slightly preceded by (315–305 Ma; Martínez Catalán, 2011a), or perhaps overlapping with (Shaw et al., 2012), formation of the CIA. The SNEO and the Iberian oroclinal structures can be interpreted therefore, as contemporaneous, antipodal, expressions of large-scale translation and drag-folding of ribbon continents along fault-bounded margins of a clockwise rotating Gondwana combined with dextral Gondwana–Laurasia shear (Klootwijk, 2009).

5. Discussion

The present model of Australia–Asia convergence-driven deformation of Australia during Middle–Late Paleozoic differs fundamentally from prevailing Paleopacific subduction-driven models. The latter are not addressed herein despite their obvious relevance to deformation of the TOS, instead focus is on potential and pitfalls of the north–south convergence-driven extrusion model.

A fundamental premise of the ASO Tectonic Extrusion Hypothesis is a uni-directional compressive force causing variably-oriented deformations, with north–south compression from Australia–Asia convergence causing, within the “compression box”, compressive deformations with tectonic grains ranging from north–south to east–west. The Middle–Late Paleozoic ASO Tectonic Extrusion Hypothesis relies substantially on observations and interpretations of Cenozoic indentation–extrusion processes within the Alpine–Himalayan belt. GPS studies show that southern Eurasian upper crustal movement vectors are progressively deflecting over up to 180° and seismic anisotropy studies show lower crustal/(uppermost lithospheric mantle?) ductile flow deflected comparably. For the widely studied India–Asia convergence/Tibetan extrusion, such systematic change in movement patterns of upper and lower crust can be attributed to gravitational and tectonic forces: gravitationally-derived deviatoric stresses from the uplifted Tibetan Plateau with flow movements guided between rigid cratonic masses; and transmission of tectonic forces that move the Indian plate by hydraulic flow processes through

ductile lower crust. Modeling of the Late Cenozoic Tibetan extrusion process has focussed largely on gravitational collapse of the uplifted Tibetan Plateau, perhaps because such a process is the more readily quantifiable. However, what is modeled is essentially a secondary–crustal uplift–process. Not the primary–crustal flow–process driven by forces that move the Indian plate northward, that is boundary forces such as basal traction, slab pull and ridge push (Liu and Yang, 2003; Bird et al., 2008; Alvarez, 2010; Copley et al., 2010; Becker and Faccenna, 2011). Hydraulic transmission of Indian tectonic forces through Asian ductile lower crust (Zhao and Morgan, 1987; Bird, 1991; Jamieson and Beaumont, 2011) deserves wider exploration and acknowledgment as a driver for Tibetan tectonic extrusion. Lower crustal hydraulic transmission is premised likewise as a prime driver for Middle–Late Paleozoic ASO-related deformation and tectonic extrusion of Australia.

It is argued here that ASO-related deformation/extrusion stems primarily from Australia–Asia convergence-driven deformation of the brittle upper crust and hydraulic deformation of the ductile lower crust. The east–west tectonic grain of the ASO derived primarily from upper crustal brittle deformation and directly reflects north–south compression from Australia–Asia convergence. The north–south tectonic grain of Quilpie Orogeny structures derived from hydraulic ductile flow through the lower crust. The east–west tectonic grain of the Olepoloko Fault Zone and part of the Kanimblan Orogeny structures probably derived from both direct upper crustal compression and hydraulic lower crustal ductile flow. Hydraulically-induced tectonic grains relate indirectly to north–south compression and more directly to local pressure gradients, eastward aligned toward the “free oceanic boundary” of the Paleopacific (Quilpie Orogeny) or southward toward the Selwyn Block (Kanimblan Orogeny). East–west compression transmitted through lower crustal ductile flow may be hard to distinguish from east–west compression driven by Paleopacific subduction, with total deformation their composite. Late Devonian and Early Carboniferous extensional deformations of the Canning and Bonaparte Basins, outside the “compression box” but synchronous with compressive deformations within the “compression box”, are tentatively attributed to lower crustal extension in regions where Australia–Asia suturing had not progressed yet, in part drawn by across-fault-leakage of compressive forces out from the “compression box”.

There is now an obvious requirement for further high quality paleomagnetic data from Australia to further detail the Devonian–Carboniferous northward excursion and to establish its reality beyond doubt. It is a sobering realization that the ASO Tectonic Extrusion Hypothesis currently hinges on one paleomagnetic interpretation for a Devonian–Carboniferous northward excursion, based solely on data from the Tamworth Belt of the SNEO obtained by a single worker (Klootwijk, 1996a, 2002, 2003, in preparation). Other workers in the SNEO, or elsewhere in Australia, have not identified this Devonian–Carboniferous northward excursion and have advocated pole path interpretations (Schmidt et al., 1990, 1993; Schmidt and Clark, 2000, 2011) substantially different from the SNEO pole path (Fig. 5; Supplementary Table 1; Klootwijk, 2009, 2010). Further substantiation of the northward excursion as interpreted from SNEO data, may have to come from other paleomagnetically prospective formations and regions, such as ignimbrites from the Yarrol Belt of the NNEO, from the Drummond Basin of the TOS, from the Kennedy Province of the Australian craton and the TOS in northern Queensland and from the Central Victorian Magmatic and Howitt Provinces of the TOS, and redbeds from Devonian and Carboniferous basins of the Australian craton and TOS.

Delineation of extent and direction of lower crustal ductile flow driven by Devonian–Carboniferous Australia–Asia convergence, has potential to open new scenarios for fluid movements and mineralization during the ASO and related orogenies. Variscan Australia–Asia convergence and Neogene convergence of Australia with the Sunda Platform

and Pacific island arcs, share comparable patterns of driving forces and locus of contact across the central New Guinean cratonic promontory. Tectonic fabrics established during Variscan convergence, such as lower crustal planar structures and crustal-penetrating detachments, are likely reactivated as easy pathways for fluid flows during Neogene convergence. Late Neogene re-heating of Cooper Basin basement (McLaren and Dunlap, 2006) may be indicative of such processes. Promising new avenues to map Variscan lower crustal ductile flow should be explored further. If the prominent negative magnetic anomalies across the Larapintine Graben and TOS represent a reverse polarity, Kiaman, remagnetization of cooled, ductilely flowed, lower crust, then its regional extent and intensity may be determinable by detailed vectorial and scalar analysis of satellite magnetic observations across altitude ranges (Thébault et al., 2010). If high seismic reflectivity of lower crust reflects sub-horizontal ductile flow rather than sub-vertical magmatic underplating (Culshaw et al., 2006; Meissner et al., 2006; Chardon et al., 2009; Dumond et al., 2010; Franěk et al., 2011), then regional extent and local direction of the flow may be mappable through reflection profiling and seismic tomography of ambient noise and teleseismic surface wave data (Li et al., 2010c; Saygin and Kennett, 2010, 2011; Yao et al., 2010; Pawlak et al., 2012) tweaked for delineating anisotropy structures within the lower crust.

6. Concluding remarks

6.1. Sources of inspiration

The concept of ASO-related tectonic extrusion sprang an afternoon in October 2006, reading David Gray and David Foster's delightful review on evolution of the Lachlan Orogen. It struck me from their figure on magnetic anomalies in the TOS (Fig. 27; Gray and Foster, 2004) that the structures resemble eastward flow lines out of central Australia. It was a small step to combine this flash of inspiration, or early onset of dementia, with smoldering ideas on Variscan Australia–Asia convergence (Klotwijk, 1995, 1996a) and an ongoing love affair with Cenozoic India–Asia convergence and Tibetan extrusion (Klotwijk et al., 1985, 1992). I downed tools on the paper at hand, a paleomagnetic study of the Rouchel Block (SNEO), and concentrated, for more years than I could have anticipated, on an extensive literature review, building and refining the tectonic extrusion concept and molding it in publishable shape. Voilà! The result is a highly provocative paper, built on a stack of heretical views, offering outrageous new ideas on several perplexing, and in part disputed, aspects of the Devonian–Carboniferous evolution of Australia and its role in Gondwana–Laurasia interaction.

It might make a heretic feel good, but how provocative really is this ASO Tectonic Extrusion Hypothesis? A summary of premises and potential new insights should indicate.

6.2. Main premises of the ASO Tectonic Extrusion Hypothesis

- (i) A newly proposed Carboniferous pole path for the SNEO resembles the KG pole path for cratonic Australia and is taken as a more detailed representation for Carboniferous movement of Australia as part of northeastern Gondwana. The KG and SNEO pole paths differ fundamentally from the SLP pole path representing the prevailing paleomagnetic view for Australia (see Klotwijk, 2009, 2010);
- (ii) The SNEO and KG pole paths detail a substantial Carboniferous northward excursion of Australia, with the cratonic promontory of central New Guinea reaching paleolatitudes of 30° to 40°N, well within the paleolatitudinal realm of the western CAOB. Northward movement of Australia/northeastern Gondwana may have started in Early Devonian, reached farthest north during the middle-late Viséan paroxysmal phase of the Variscan

Orogeny, and reversed into very fast southward movement toward polar latitudes;

- (iii) Convergence and collision of eastern Gondwana with southern Laurasia is advanced as the common cause of Variscan orogenic deformation throughout the CAOB and Australia, as expressed by the Alice Springs, Quilpie and Kanimblan Orogenies;
- (iv) Convergence-related compressional deformation was confined to a “compression box”, extending southward from the central New Guinean cratonic promontory, including cratonic Australia and the TOS, but excluding the NEO to its east and also westernmost Australia;
- (v) Tectonic extrusion was facilitated by heated, weakened, crust within the Larapintine Graben and the TOS and the “free oceanic boundary” of the Paleopacific. Extrusion progressed through ductile flow, and probably partial melt, in the lower crust, with hydraulic re-alignment of originally north–south compressional forces along east–west regional pressure gradients in the eastern Thomson Orogen and NNEO.

6.3. New insights into Australia's Middle–Late Paleozoic tectonic evolution

- (i) A realistic driver for Australia-wide north–south compressional deformation of ASO-vintage has been wanting. Prevailing models figuring westward to highly oblique Paleopacific subduction seemingly lack reality. Australia's paleomagnetically-based northward excursion indicating convergence with southern Laurasia mainly during Early Carboniferous, offers tectonic extrusion as a plausible explanation for contemporaneous north–south deformation of the ASO, east–west deformation of the Quilpie Orogeny, and a north–south deformation component of the Kanimblan Orogeny. Different tectonic grains of the ASO (E–W), Quilpie Orogeny (N–S) and Kanimblan Orogeny (part E–W) represent different effects of Australia–Asia convergence on the brittle upper crust and ductile lower crust. In the upper crust, direct transmission of north–south compression primarily led to the east–west tectonic grain of the ASO. In the lower crust, hydraulic re-alignment of north–south compression along prevailing east–west pressure gradients near the “free boundary” of the Paleopacific, led to the north–south tectonic grain of the Quilpie Orogeny, with further flow re-alignment contributing to the east–west tectonic grain component of the Kanimblan Orogeny. Variscan deformation of Australia is the composite effect of far-field forces transmitted directly through the brittle upper crust and hydraulically through the ductile lower crust, resulting in widely varying, interrelated, tectonic grains;
- (ii) Coincidence of Devonian–Carboniferous compression in ASO-affected regions of the Australian craton with extension in the Canning and Bonaparte Basins of western Australia has been a baffling observation. The concept of a “compression box”, extending southward from the cratonic promontory of central New Guinea, offers a simple explanation for contemporaneous occurrence of compressional deformation within the box, far lesser deformation outside the box in the NEO, and extension in western Australia. Variscan Australia–Asia suturing progressed beyond the central New Guinean cratonic promontory during the mid Viséan collisional climax, with compressional deformation expanding into western Australia, widening the “compression-box”;
- (iii) Seismic tomography studies of Australian uppermost lithospheric mantle/lower crust consistently show continental-like velocities for the internal TOS. Such velocities are not expected for a TOS developed from oceanic basement, but are herein interpreted to represent cratonic influx, tectonically extruded from the craton through lower crustal ductile flow. East–west fanning out patterns of azimuthal anisotropy in central and eastern Australia reflect flow patterns of lower crust extruded from the Larapintine Graben;

- (iv) Prominent negative magnetic anomalies across the internal TOS and Larapintine Graben reflect reverse polarity overprinting of deeper crust, of post-tectonic extrusion, Kiaman, origin. As an upper crustal corollary to such deeper crustal magnetic overprinting, widespread presence of Devonian and Early Carboniferous paleomagnetic overprints throughout the TOS and Larapintine Graben represents syn-tectonic-extrusion, mainly pre-Kiaman, hydrothermal imprints derived from lower crustal ductile flow and partial melt. This dual pattern of pervasive deeper crustal Late Paleozoic, Kiaman, remagnetization and upper crustal Middle–Late Paleozoic paleomagnetic overprinting resembles well-established remagnetization patterns for European Variscan massifs;
- (v) Further detail on the extent of lower crustal flow may come from mapping of seismically highly reflective lower crust, assuming that its planar anisotropy reflects horizontal ductile flow rather than vertical magmatic underplating, and from mapping of magnetic overprinting of deeper and uppermost crust;
- (vi) The Namurian lacunae, characteristic for most of cratonic Australia, are likely to reflect tectonically- and/or thermally-induced uplift within the “compression box” and consequent surficial erosional transport from the thermally elevated “compression box” into tectonically less affected NEO and western Australian basins. Thermal expansion induced by tectonically extruded, heated, lower crust led to Late Carboniferous development of the Kanimblan Highlands (TOS), with latest Carboniferous and Permian thermal relaxation leading to their demise. Enhanced middle to Late Carboniferous lower crustal heat flux and resulting temporal change in the paleo-geothermal gradient are more likely to explain the fission track and low-temperature isotopic disturbance record than the disputed requirement for extensive burial and denudation;
- (vii) Eastward tectonic extrusion of the Thomson Orogen and NNEO, subsequent development of the EARS, and latest Carboniferous–Early Permian clockwise rotation of Australia/Gondwana, offer a new model for development of the Texas–Coffs Harbour and Manning Oroclines of the SNEO as mega-dragfolds of a ribbon continent. Clockwise rotation of cratonic Australia led to northward telescoping of an unpinning SNEO against the eastward-extended buttress of the NNEO, with telescoping facilitated by detachment of the SNEO continental ribbon from the Lachlan Orogen along the developing EARS. Development of the SNEO oroclinal is mirrored, in time, location and cause, at its Gondwana antipode by oroclinal development across European Variscan massifs, Iberian in particular.

Supplementary data to this article can be found online at <http://dx.doi.org/10.1016/j.gr.2012.10.007>.

Acknowledgments

I cordially thank Russell Korsch (Geoscience Australia), John Laurie (GA) and Peter Jones (RSES) for stratigraphic guidance, Lothar Ratschbacher (Technische Universität Bergakademie Freiberg) for Fig. 9A, Chris Fitzgerald (GA) for Fig. 11A, Peter Milligan (GA) for Figs. 21H, I and 24F, Prame Chopra (Earthinsight) for Fig. 24C, Brad Pillans (RSES) for access to the Black Mountain Paleomagnetic Laboratory, and Richard Arculus, David Ellis, Brian Kennett and Andrew Roberts (RSES) for offering a visiting fellowship to a refugee in need. I kindly thank the two anonymous reviewers for their positive comments that have been most useful to improve presentation and focussing of the paper. Last but not least I wish to thank the authors of the many excellent papers that I have read and re-read over the six years gestation of this review. Your information and insights have made for many stimulating hours of reading and have been

most helpful in guiding my thinking and in formulating this highly unconventional hypothesis. Regrettably I could cite and reference only a subset of papers worthy of quotation. Of course I am solely to blame for heretical views.

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Glossary

- ANU: The Australian National University
 ASO: Alice Springs Orogeny
 CAA: Cantabrian–Asturian Arc
 CAOB: Central Asian Orogenic Belt
 CIA: Central Iberian Arc
 CNPS: Cretaceous Normal Polarity Superchron
 CRFZ: Clarke River Fault Zone
 EARS: East Australian Rift System
 GA: Geoscience Australia
 GPE: Gravitational Potential Energy
 GPS: Global Positioning System
 HHGG: High Heat Generating Granite
 KG: Klotwijk and Giddings pole path
 NEO: New England Orogen
 NNEO: Northern New England Orogen
 PCRS: Permo-Carboniferous Reverse Superchron
 RSES: Research School of Earth Sciences
 SHRIMP: Sensitive High Resolution Ion Microprobe
 SLP: Schmidt, Li and Powell pole path
 SNEO: Southern New England Orogen
 TOS: Tasman Orogenic System

Chris Klootwijk enjoys life as a Visiting Fellow at the Research School of Earth Sciences, The Australian National University. He studied geology (BSc Cum Laude, MSc Cum Laude) and geophysics (MSc Cum Laude) at the University of Utrecht and obtained there a PhD for a paleomagnetic study on the drift of India. He carried out further studies on the India–Asia collision at the Australian National University, Centre National de la Recherche Scientifique and the Université de Paris 7, before returning to Australia to oversee the paleomagnetism group of the Bureau of Mineral Resources/Australian Geological Survey Organisation. For the past decade at ANU, between walking with his German Shepherd dog over the beautiful Canberra hills and increasingly sedate kayaking, he is trying to shed some light on the tectonic evolution of Australia while keeping abreast of exciting developments on Himalayan–Tibetan tectonics.

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